

The 1976 Standard Atmosphere Above 86-km Altitude

Recommendations of
Task Group II to COESA

National Aeronautics and
Space Administration



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**Recommendations of
Task Group II to COESA**

Edited by

R. A. Minzner
Goddard Space Flight Center

Prepared by NASA Goddard Space Flight Center



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SYMBOLS

a	A constant coefficient, -19.9429 km, used to specify the elliptical segment of the temperature-height profile between 91 and 110 km, as defined by equation 2a
a_i	A species-dependent coefficient ($\text{m}^{-1} \text{s}^{-1}$) listed in table 4 and used in equation 14 to define D_i for the gases O, O ₂ , Ar, and He
A	A constant coefficient, -76.3232 K, used in equation 2a to specify the elliptical segment of the temperature-height profile between 91 and 110 km
b	A dimensionless subscript representing a set of integers as listed in table 3
b_i	A dimensionless, species-dependent exponent listed in table 4, and, along with a_i , used in equation 14 to define D_i
D_i	The height-dependent, species-dependent, molecular-diffusion coefficient ($\text{m}^2 \text{s}^{-1}$) defined in equation 14 for O, O ₂ , Ar, and He, and in equation 30 for atomic hydrogen
$F(Z)$	The designation of a particular functional dependence upon Z (m^{-1}) as defined by equation 10
g	The height-dependent, 45°-latitude accelerations of gravity ($g(Z)$) (m s^{-2}), implicit in H_p and H_i as defined by equations 5a and 5b, respectively, and related to g_0 in equation 5c
g_0	The standard, 45°-latitude, sea-level value of g , 9.80665 m s^{-2}
H	Geopotential height (expressed in units of a standard geopotential meter (m'), which in this document is called simply a geopotential meter) used as the argument for all tables in the height range 0 to 84.852 km
H_i	The height-dependent scale height (m) of the i th gas species, defined in equation 5b
H_p	The height-dependent pressure scale height (m) of the mixture of gases comprising the atmosphere, defined in equation 5a
i	A dimensionless subscript designating the i th member of a set of gas species
K	The height-dependent, eddy-diffusion coefficient ($\text{m}^2 \text{s}^{-1}$) referred to in equation 5, and defined in equations 11, 12, and 13
L	The height-dependent mean free path (m) defined in equation 24
L_K	The gradient (K/km) of T with respect to Z , that is, dT/dZ

L'_M	The gradient (K/km') of T_M with respect to H , that is, dT_M/dH
M	The height-dependent mean molecular weight (kg kmol ⁻¹) defined in equation 19 for a mixture of gases constituting the atmosphere
M_i	The molecular weight (kg kmol ⁻¹) of the i th gas species
M_0	The standard, sea-level value of M , 28.9644 kg kmol ⁻¹
n_i	The height-dependent number density (m ⁻³) of the i th gas species, which in the case of N_2 is related to T and g in equation 7, while for O , O_2 , Ar , and He , is related to T , D_i , and K in equation 9
N	The height-dependent total number density (m ⁻³) defined by equation 17 for a mixture of neutral atmospheric gas particles
P	The height-dependent atmospheric pressure (N m ⁻² or equivalently Pa, both equal to 10 ² mbar)
q_i	One of six species-dependent dimensional constants, that is, q_i , Q_i , u_i , U_i , w_i , and W_i , all listed with dimensions in table 5, and all used in equation 16, an integrable expression for $v_i/(D_i + K)$
r	The effective Earth's radius, 6356.766 km, used in equation 5c to compute the height dependence of g for 45° North latitude
R^*	The universal gas constant, 8.31432×10^3 N m K ⁻¹ kmol ⁻¹
T	The height-dependent kinetic temperature in the Kelvin scale (K), defined as a function of Z in equations 1a through 4b
T_b	Temperatures defined in table 3
T_c	A constant, 263.1905 K, the temperature coordinate of the center of the ellipse, defining $t(Z)$ between 91 and 110 km in equation 2a
T_M	The height-dependent molecular-scale temperature (K), equal to kinetic temperature T times the ratio M/M_0 , and listed in table 1
T_r	The temperature, 999.2356 K at $Z_r = 500$ km
T_∞	The exospheric temperature taken to be 1000 K, and used in equation 4a
v_i	The flow velocity of the i th gas species, as used in equations 5, 9, and 16
V	The height-dependent mean particle speed (m s ⁻¹) defined in equation 22
Z	Geometric height (m) used as the argument of the tables and graphs at heights above 86 km
Z_b	Specific reference heights as defined in table 3

Z_c	The height coordinate, 91 km, of the center of the ellipse defining $T(Z)$ between $Z_1 = 91$ and $Z_2 = 110$ km
Z_r	A specific reference height which is 500 km in equation 26
α	The dimensionless, thermal-diffusion factor which has the value -0.25 for atomic hydrogen in equations 25 and 26
α_i	The dimensionless, species-dependent, thermal-diffusion factor which has the value -0.4 for He, and zero for O, O_2 , and Ar in equations 5 and 10
ξ	The geopotential (km') of various heights, Z , relative to that at a reference level $Z_3 = 120$ km, as used in equation 4a
λ	A constant coefficient, 0.01875 ($1/km'$) in equation 4a defining $T(Z)$ between 120 km and 1000 km
ν	The height-dependent mean collision frequency (s^{-1}) defined in equation 23
ζ	The value of an unspecified series ($m^{-3} s^{-1}$) representing production and loss terms in an expression of continuity in equation 8
ρ	The mass density of air ($kg m^{-3}$) related to number density in equation 18
σ	The effective mean collision diameter, 3.65×10^{10} (m), used in the expression for ν and implicitly in the expression for L
τ	A dimensionless coefficient representing the reduced height of the atomic hydrogen relative to a reference height $Z_r = 500$ km
ϕ	The vertical flux ($m^{-2} s^{-1}$) of atomic hydrogen

INTRODUCTION

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A meeting of the Committee for the Extension to the U.S. Standard Atmosphere (COESA) was held in Boston on September 13, 14, and 15, 1971. At this meeting it was agreed to review the current state of knowledge of the Earth's atmosphere with the objective of possibly revising the *U.S. Standard Atmosphere, 1962* (COESA, 1962) above the altitude of about 50 km, and of extending the upper limit of the tabulations of a revised standard atmosphere to 1000 km. In order to distribute the workload of review and revision, assignments were made to various task groups in accordance with particular height regions of the atmosphere.

Several schemes have been used to divide the vertical extent of the atmosphere into a number of regions. One of these is based upon the major features of the temperature-height profile. Another is based upon major features in the height distribution of the constituent gases of the atmosphere. A third is based upon the principal techniques used for measuring the thermodynamic properties of the atmosphere (temperature, density, and pressure), while a fourth is dependent upon the amount of observational data available for constructing and testing any proposed model. Based primarily upon the latter two schemes, that is, measuring technique and quantity of data, COESA divided the atmosphere into three overlapping regions, and assigned a particular task group to concentrate on each of these three regions:

Task Group I	50 to 100 km
Task Group II	80 to 200 km
Task Group III	140 to 1000 km

The height region for Task Group I, 50 to 100 km, is a region in which atmospheric measurements of temperature, density, and pressure are made almost exclusively with rocket-borne instruments. These instruments have served to develop an extensive set of thermodynamic data, particularly below 90 km. The height region for Task Group III, that is, 140 to 1000 km, is one in which the thermodynamic properties are determined almost exclusively from satellite-related observations. A vast amount of data, particularly mass-density data, has been accumulated for this height region.

The height region for Task Group II (80 to 200 km, but primarily the 65 km between 86 and 150 km) is a region for which there is only a very limited amount of atmospheric data from rocket soundings, and almost none from satellite observations. Furthermore, no unique vehicle or observational technique has, to date, been developed for efficient observation of the thermodynamic and photochemical properties of this region of the Earth's

atmosphere. Consequently, only a relatively small amount of data for this height region was available for consideration, and the job of Task Group II became primarily one of defining mutually compatible boundary conditions, and of defining a series of analytical functions to bridge the gap between models which were appropriate to the other two regions.

Task Group II, the membership of which is listed on page iii, met on four different occasions:

February 22, 1972	GSFC	Greenbelt, Maryland
April 19, 1972	Sheraton Park Hotel	Washington, D.C.
November 3, 1972	GSFC	Greenbelt, Maryland
December 8, 1972	AFCRL	Bedford, Massachusetts

with K. S. W. Champion, C. A. Reber, and R. A. Minzner in attendance at each of these meetings, and with other members absent from one or more meetings.

Two basic proposals regarding the analytical functions were submitted to the committee—one by G. R. Swenson and one by C. A. Reber. The first two meetings were devoted to the discussion of these proposals, and successive revisions were made between meetings. By the third meeting, some of the differences between these two proposals had been resolved, and G. R. Swenson withdrew his proposal in favor of the successively revised Reber model. At this meeting, empirical number-density concentrations of various atmospheric species for the 150-km height level were agreed upon by the committee. Some revisions to eddy-diffusion coefficients were also suggested. To achieve the recommended N_2 densities at 150 km, as well as to incorporate some recommended revisions in Reber's temperature-height profile in the 100- to 115-km region, it became necessary, through coordination with Task Group I, to lower the temperature in the 86- to 90-km region from that originally suggested by Task Group I.

At the December 1972 meeting, the revised version of the Reber model was reviewed, and further modifications were suggested. The committee agreed to adopt the Reber model of that date, with a few additional modifications, for submission to COESA.

The original edition of this report was prepared and submitted to COESA at its February 1973 meeting (in conjunction with the annual meeting of the American Meteorological Society in St. Petersburg, Florida). The boundary conditions, as well as the general aspects of the Task Group II model, were adopted at this meeting. The final tables, however, were to await the inclusion of revised eddy-diffusion coefficients which were later supplied by the Air Force Cambridge Research Laboratory (AFCRL). These values were received in February 1974 when a function approximating these new coefficients was prepared. These new tables were distributed at a joint meeting of the Working Group of COESA and Task Group IV, which was established to coordinate the work of the other three task groups. At this meeting, held at Bedford, Massachusetts on March 27, 1974, Task Group II was asked to expand its tables in two areas: the first was to include hydrogen densities in

accordance with calculations by Capt. Forbes of AFCRL (see Appendix A); the second was to extend the height range of the entire set of tabulations to 1000 km, thereby effectively producing the tables for Task Group III. This extension was accomplished with relative simplicity because the analytical function of the upper segment of the model developed by Task Group II is identically the function used by Task Group III for the height region up to 1000 km.

The temperature-height and density-height profiles of the coordinated atmospheric model finally agreed upon by Task Group IV and later adopted by COESA are presented in Appendix B. This appendix also presents comparable graphs of the *U.S. Standard Atmosphere, 1962* (COESA, 1962) and of many earlier standard atmospheres which strongly influenced the development of the currently agreed upon model. These graphs permit a comparison of the several models and help explain the history of the contribution of each of these models to the development of the *U.S. Standard Atmosphere, 1976* (COESA, 1976).

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LOWER BOUNDARY CONDITIONS

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Because of the necessity for continuity with the *U.S. Standard Atmosphere, 1962* (COESA, 1962) at a height of 50 km, it was necessary for the several task groups to generate their respective models sequentially. Task Group II was required, therefore, to use the temperature and density of the model of Task Group I, at some point in the overlap region between 80 and 100 km, as the lower boundary values for its model.

The purpose of the overlap in the height regions assigned to each of the task groups was to allow needed flexibility in the generation and matching of the successive models. Such flexibility was particularly necessary in the height regions of overlap between the model of Task Group I (hereafter referred to as Model I) and the model of Task Group II (hereafter referred to as Model II). This strong need for flexibility stems, in part, from the fact that some fundamental differences exist in the units of the defining properties used in the respective models.

Model I, in conformity with the constraints of the existing U.S. Standard Atmosphere (COESA, 1962), is defined in terms of geopotential height H (measured in geopotential kilometers (km')), and L_M the height gradient of molecular-scale temperature, T_M . The quantity L_M is specifically defined as dT_M/dH where T_M is a special kind of temperature which is equal to kinetic temperature times the ratio of sea-level value of molecular weight, M_0 , to the ambient value of molecular weight, M . Both geopotential height and molecular-scale temperature are discussed elsewhere (Minzner et al., 1976). Furthermore, Model I is characterized by a set of particular constant values of L_M , each applying respectively to one of a number of specific layers. This model is presented in skeleton form in table 1, with defining quantities underlined. In contrast, the defining parameters of Model II are kinetic temperature, T , and geometric height, Z (measured in geometric kilometers (km)), with no specific requirements for defining any particular layer in terms of a constant value of the gradient of kinetic temperature with respect to height, that is, dT/dZ need not necessarily be constant over any height region of Model II.

Since a constant value of dT_M/dH is exactly equal to a constant value of dT/dZ in a vertical isothermal layer, within which there is no variation of mean molecular weight, it would be most desirable to design the transition from the T_M -versus- H definition of Model I to the T -versus- Z definition of Model II within such an isothermal layer. Individual atmospheric

Table 1
A Skeleton Version of Model I

Z km	H km'	T _M K	L _M ' K/km'	P mbar	ρ kg m ⁻³	N m ⁻³
0	<u>0.000</u>	<u>288.150</u>	-6.5	<u>1.013250 × 10³</u>	1.224999 × 10 ⁰	2.546972 × 10 ²⁵
	<u>11.000</u>	216.650	<u>0.0</u>	2.263206 × 10 ²	3.639178 × 10 ⁻¹	7.566441 × 10 ²⁴
	<u>20.000</u>	216.650	<u>+1.0</u>	5.474889 × 10 ¹	8.803480 × 10 ⁻²	1.830386 × 10 ²⁴
	<u>32.000</u>	228.650	<u>+2.8</u>	8.680187 × 10 ⁰	1.322500 × 10 ⁻²	2.749692 × 10 ²³
	<u>47.000</u>	270.650	<u>0.0</u>	1.109063 × 10 ⁰	1.427532 × 10 ⁻³	2.968072 × 10 ²²
	<u>51.000</u>	270.650	<u>-2.8</u>	6.693887 × 10 ⁻¹	8.616049 × 10 ⁻⁴	1.791416 × 10 ²²
	<u>71.000</u>	214.650	<u>-2.0</u>	3.956420 × 10 ⁻²	6.421099 × 10 ⁻⁵	1.335051 × 10 ²¹
<u>86</u>	84.852	186.946		3.733836 × 10 ⁻³	6.957879 × 10 ⁻⁶	

soundings, when averaged, show the relatively sharp mesopause to be smeared over an extended height range leading to an isothermal layer in the model. This layer appears, at first, to provide the desired transition region. Unfortunately, however, the conditions associated with molecular oxygen dissociation, accounted for in Model II, propagate downward to a small extent through and below this region. Thus, while dT/dZ might be defined to be zero, dM/dZ has a small nonzero value which increases with increasing height, and in this isothermal layer, dT_M/dH is not exactly equal to dT/dZ . Consequently, the optimum transition height from Model I to Model II is not located within the isothermal layer, but, rather, is located at its base, where dM/dZ has its smallest value for this layer.

Because the integration process used in the development of Model II is based on height increments of exactly 1 km, it is most desirable that the integration would begin at an integer kilometer value. Consequently, the base of the above-mentioned isothermal layer has been designated to coincide exactly with an integer geometric-kilometer value.

The constraints of observed temperature-height profiles in the 70- to 85-km region and in the 110- to 120-km region, as well as the constraint of observed densities in the 150-km region, indicated the need for a temperature of about 187 K in the isothermal layer. Along the constant gradient of -2 K/km', extending upward from 71 km' in Model I, the integer geometric-kilometer value most nearly associated with a temperature of 187 K is 86 km (or 84.852 km'). At this height, $T_M = 186.946$ K, and the value of M is sufficiently close to the value of M_0 so that T differs from T_M by less than 0.1 K. Thus, 84.852 km' (or identically 86 km) was finally selected as the most desirable height for the interface between Model I and Model II.

The value, $T_M = 186.946$ K at $H = 84.852$ km', stems from the value of $T_M = 214.65$ K at $H = 71$ km' and the gradient of -2 K/km' extending upward for 13.852 km' from 71 km'. Thus, at 84.852 km', $T_M = 214.65$ K + $[(-2$ K/km') $\times 13.852$ km'] = 186.946 K.

The composition of the atmosphere between 86 and 150 km in Model II includes molecular nitrogen, N_2 ; molecular oxygen, O_2 ; argon, Ar; helium, He; and atomic oxygen, O. At

150 km, atomic hydrogen, H, is introduced. The first three constituents are those having the greatest fractional concentrations between sea level and 86 km, while the remainder have very small fractional concentrations in this height region. The model does not include carbon dioxide, CO₂; neon, Ne; krypton, Kr; xenon, Xe; methane, CH₄; and molecular hydrogen, H₂; all of which contribute somewhat to the standard six-significant-figure value of the sea-level mean molecular weight, 28.9644 kg kmol⁻¹. Because of the adopted composition in Model II, the 86-km mean molecular weight differs from the sea-level value and is seen to be 28.95221 kg kmol⁻¹ (Minzner et al., 1976). This value, along with the value of M₀ and the value of T_M at 86 km, implies an 86-km kinetic temperature, T = 186.87631 K through the relationship $T = T_M (M/M_0)$.

The continuity of T_M, or the ratio T_M/M₀, across the 86-km boundary between Model I and Model II implies continuity in those atmospheric parameters whose height variation depends upon the ratio T/M or its reciprocal. This continuity exists because T/M is equal to T_M/M₀, a ratio which at 86 km has a value 6.454337 K kmol kg⁻¹. Thus, the values of pressure, P; density, ρ; pressure scale height, H_p; and mean particle speed, V—all of which depend upon the ratio T/M—will each be continuous across this boundary.

The common boundary values of P and ρ can be found in table 1. The boundary values of H_p and V are 5621.212 m and 369.6658 m s⁻¹, respectively. The values of mean free path, L, mean collision frequency, ν, and total number density, N, however, cannot be computed in terms of the ratio T/M or T_M/M₀ alone, but rather, require the additional information of the value of T or M independent of the ratio T/M. Thus, Model I, which does not independently define values of T or M at heights where M departs from M₀, cannot accurately specify values of L, ν, and N at these heights. It is assumed that this departure of M from M₀ takes place smoothly between 79 km' and 84.52 km' (that is, between 80 and 86 km). The values of L, ν, and N have been computed for 86 km on the basis of T = 186.8673 K, M = 28.9522, the standard effective collision cross-section diameter σ = 3.65 × 10⁻¹⁰ m, and the particular adopted values of R* and N_A, that is, 8.31432 × 10³ N m K⁻¹ kmol⁻¹ and 6.022169 × 10²⁶ kmol⁻¹, respectively. These boundary values of L, ν, and N, as well as those of properties already discussed, were all computed from equations in the *U.S. Standard Atmosphere, 1962* (COESA, 1962) and the values are summarized below:

T _M	=	186.9460 K		
M ₀	=	28.9644 kg kmol ⁻¹		
T	=	186.8637 K		
M	=	28.9522 kg kmol ⁻¹		
P	=	0.3733836* N m ⁻²	log ₁₀ P	= - 0.42784476
ρ	=	6.957879 × 10 ⁻⁶ kg m ⁻³	log ₁₀ ρ	= - 5.15752313
H _p	=	5621.212 m		

$$V = 369.6658 \text{ m s}^{-1}$$

$$\nu = 3.166708 \times 10^4 \text{ s}^{-1} \quad \log_{10} \nu = + 4.50060802$$

$$L = 1.167350 \times 10^{-2} \text{ m} \quad \log_{10} L = - 1.93279891$$

$$N = 1.447265 \times 10^{20} \text{ m}^{-3} \quad \log_{10} N = +20.16054806$$

*100 N m⁻² = 1 mbar. Thus, at 86 km we may also specify

$$P = 3.733836 \times 10^{-3} \text{ mbar} \quad \log_{10} P = - 2.42784476$$

PHILOSOPHY AND CONSTRAINTS OF MODEL II

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In addition to conforming to the basic guidelines already mentioned, the mathematical expressions defining the temperature and number-density profiles as a function of altitude are formulated to make the model useful as a theoretical tool over its entire altitude range. A number of implications follow from this philosophy:

- The temperature is expressible as a smooth mathematical function of geometric altitude, with a continuous first derivative. An exponential profile (Bates, 1959) is used in the middle and upper thermosphere, as this form is well known, widely used, and permits the utilization of the techniques of Bates (1959) and Walker (1965) for analytically representing upper-atmosphere number densities.
- The functions representing the temperature-height profile are readily adjustable to allow the model to be adapted to various data sets.
- The functions relating number densities to altitude are physically meaningful and analytically expressible.

Furthermore, the temperatures and gas-species number densities of the model are consistent with the limited observational data from a variety of sources.

- At the lower altitude boundary of Model II, the temperature, pressure, mass density, scale height, and mean particle speed match those of Model I, a condition which is somewhat complicated by the fact that Model I is defined in terms of geopotential altitude and gradients of molecular-scale temperature, while Model II uses geometric altitude and kinetic temperature.
- In the height region between the lower boundary of Model II and about 130 km, the temperature and mass-density profiles match the available data, which come largely from rocket-borne pitot-tube measurements, falling-sphere measurements, and Thomson incoherent-scatter measurements. (However, the average value of the N_2 density-height profile above 150 km is fairly well established, and the number densities at these heights have a large influence on the choice of temperature profiles in the region below this altitude. See the section, Height Profiles of Temperature and N_2 Number Density.)
- At 150 km, the composition matches the recommendations of Task Group II. These recommendations, which are the values shown in table 2 for 150 km, and which are also discussed in the section on Composition, are those which were accepted by COESA at its meeting in St. Petersburg during February 1973.

Table 2
Number Densities of Five Species, as well as Mass Density and
Mean Molecular Weight at Selected Heights

Gas	86 km	120 km	150 km	450 km
N ₂	$1.12979 \times 10^{20} \text{ m}^{-3}$	$3.7224 \times 10^{17} \text{ m}^{-3}$	$3.1211 \times 10^{16} \text{ m}^{-3}$	$1.0855 \times 10^{12} \text{ m}^{-3}$
O (atomic)	$8.60000 \times 10^{16} \text{ m}^{-3}$	$9.2746 \times 10^{16} \text{ m}^{-3}$	$1.7800 \times 10^{16} \text{ m}^{-3}$	$4.1636 \times 10^{13} \text{ m}^{-3}$
O ₂	$3.03090 \times 10^{19} \text{ m}^{-3}$	$4.3949 \times 10^{16} \text{ m}^{-3}$	$2.7500 \times 10^{15} \text{ m}^{-3}$	$2.3676 \times 10^{10} \text{ m}^{-3}$
He	$7.58173 \times 10^{14} \text{ m}^{-3}$	$3.8878 \times 10^{13} \text{ m}^{-3}$	$2.1058 \times 10^{13} \text{ m}^{-3}$	$3.9478 \times 10^{12} \text{ m}^{-3}$
Ar	$1.35140 \times 10^{18} \text{ m}^{-3}$	$1.6361 \times 10^{15} \text{ m}^{-3}$	$5.0000 \times 10^{13} \text{ m}^{-3}$	$2.6583 \times 10^7 \text{ m}^{-3}$
H (atomic)	—	—	$3.7541 \times 10^{11} \text{ m}^{-3}$	$8.4429 \times 10^{10} \text{ m}^{-3}$
ρ	$6.95788 \times 10^{-6} \text{ kg m}^{-3}$	$2.221 \times 10^{-8} \text{ kg m}^{-3}$	$2.075 \times 10^{-9} \text{ kg m}^{-3}$	$1.184 \times 10^{-12} \text{ kg m}^{-3}$
M	$28.95220 \text{ kg kmol}^{-1}$	$26.204 \text{ kg kmol}^{-1}$	$24.102 \text{ kg kmol}^{-1}$	$15.247 \text{ kg kmol}^{-1}$

- An extrapolation of the model temperature into the exosphere yields values asymptotic to 1000 K in accordance with a COESA decision of September 1971.
- In the height region of the thermosphere, the neutral composition of the model is in agreement with observational data. These data come primarily from the quadrupole mass spectrometer on the Orbiting Geophysical Observatory (OGO)-6 satellite (for example, Hedin et al., 1972). Since these data are referenced primarily to an altitude of 450 km, it was decided to extend the definition of Model II to at least that altitude to permit the coefficients of the exponential segment of the model to be chosen in accordance with this large and unique data set. The values of N₂, O₂, O, and He for 450 km in table 2 represent the OGO-6 data after its adjustment to 45°N latitude and to an exospheric temperature of 1000 K. These adjusted OGO-6 data are the ones to which Model II has been matched.
- At altitudes above about 150 km, the total density and the pressure scale height are consistent with the large body of data determined from satellite drag.
- The number densities, eddy-diffusion coefficient, and flux terms of the steady-state representation of Model II are consistent with the results of the theory developed by Keneshea and Zimmerman in their model involving dynamic and photochemical transport (Minzner et al., 1976). This theory incorporates eddy-diffusion coefficients based principally on data gathered during the ALADDIN I experiment (Philbrick et al., 1973), an experiment for measuring atmospheric layering and density distribution of ions and neutrals.

While the parameters and the height profiles of the various atmospheric properties used in or developed by these calculations represent a kind of steady-state condition, the true state of the atmosphere is dynamic, and the computed model described herein is only an approximation to actual conditions. Some examples of the dynamic aspects of the atmosphere include the variability of the following properties: the temperature profile which generally exhibits wave-like structure, the atomic oxygen profile which may be shown to be extremely time dependent with a significant diurnal component, and the helium profile which has an annual component. For the purposes of this segment of the *U.S. Standard Atmosphere, 1976* (COESA, 1976), however, the choices of boundary conditions and anchor points were made with the objective of obtaining the best average representation of the dynamic atmosphere.

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DEFINITION OF MODEL II

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Silver Spring, Maryland*

TEMPERATURE-HEIGHT PROFILE

The constraints imposed upon Model II by the data and the philosophy of the model led to the adoption of a temperature-height profile specified in four consecutive segments. The first segment represents an isothermal layer between 86 and 91 km, while the third represents a layer of linearly increasing temperatures between particular values of temperature at 110 and 120 km, consistent with a gradient dT/dZ equal to 12 K/km. The second segment is a portion of an ellipse designed so that the end-point temperatures and gradients exactly match those of the adjacent segments. The fourth segment is an exponential designed so that T , in approaching 1000 K asymptotically at $Z = \infty$, is within 1° of 1000 K at 500 km, and so that T and dT/dZ (with the latter designated by the symbol L_K) are both continuous between the third and fourth segments at their 120-km junction height.

The end-point values of Z , T , and L_K , that is, Z_b , T_b , and $L_{K,b}$ respectively, and the related values of b , the subscript associated with each end-point or junction height, are listed in table 3. The subscript values begin with 7 in keeping with the fact that level 7 is simultaneously the top of the seventh segment of Model I (below 86 km) and the base of the first segment of Model II.

A graph of that portion of the temperature-height profile between 86 and 140 km is shown in figure 1.

The four segments of the temperature-height profile of Model II, and the associated gradients are mathematically defined below. That segment of T versus Z between $Z_7 = 86$ km and $Z_8 = 91$ km is represented by the special case of the linear equation in which the gradient is zero, so that

$$T = T_7 \tag{1a}$$

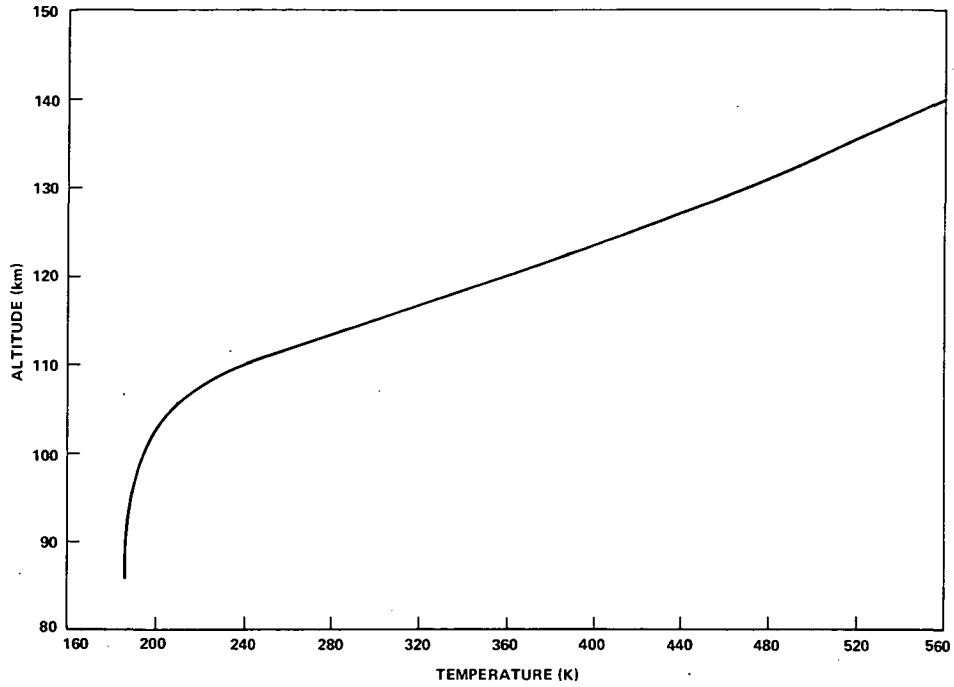


Figure 1. Atmospheric temperature versus altitude, 86 to 140 km.

Table 3
End-point Values of the Four Segments of the Model II
Temperature-height Profile

Subscript b	Height Z_b (km)	Temperature T_b (K)	Gradient $L_{K,b}$ (K/km)
7	86	186.8673	0.0
8	91	186.8673	0.0
9	110	240.0	12.0
10	120	360.0	12.0
∞	∞	1000.0	0.0

where

$$T_7 = 186.8673 \text{ K},$$

and

$$\frac{dT}{dZ} = 0.0 \quad (1b)$$

That segment of T versus Z between $Z_8 = 91$ km and $Z_9 = 110$ km is mathematically specified by a segment of an ellipse (Minzner et al., 1976), such that

$$T(Z) = T_c + A \left[1 - \left(\frac{Z - Z_8}{a} \right)^2 \right]^{1/2} \quad (2a)$$

where

$$T_c = 263.1905 \text{ K},$$

$$A = -76.3232 \text{ K},$$

$$Z_8 = 91 \text{ km},$$

$$a = -19.9429 \text{ km},$$

and

$$\frac{dT}{dZ} = \frac{-A}{a} \left(\frac{Z - Z_8}{a} \right) \left[1 - \left(\frac{Z - Z_8}{a} \right)^2 \right]^{-1/2} \quad (2b)$$

That segment of T versus Z between $Z_9 = 110$ km and $Z_{10} = 120$ km is represented by the linear function

$$T = T_9 + L_{K,9} (Z - Z_9) \quad (3a)$$

where

$$T_9 = 240 \text{ K},$$

$$L_{K,9} = 12 \text{ K/km},$$

$$Z_9 = 110 \text{ km},$$

and

$$\frac{dT}{dZ} = L_{K,9} \quad (3b)$$

Finally, the segment of T versus Z between $Z_{10} = 120$ km and $Z_\infty = \infty$, is a particular exponential function (Walker, 1965)

$$T = T_\infty - (T_\infty - T_{10}) \cdot \exp(-\lambda \xi) \quad (4a)$$

where

$$T_{\infty} = 1000 \text{ K},$$

$$T_{10} = 360 \text{ K},$$

$$\lambda = [1/(T_{\infty} - T_{10})] \cdot L_{K,10} = 0.01875 \text{ (1/m')},$$

$$\xi = (Z - Z_{10}) (r + Z_{10}) / (r + Z)$$

in which expressions

$$L_{K,10} = 12 \text{ K/km}$$

$$Z_{10} = 120 \text{ km},$$

$$r = 6356.766 \text{ km, the effective Earth's radius at } 45^{\circ}\text{N latitude.}$$

The derivative of equation 4a yields

$$\frac{dT}{dZ} = \lambda (T_{\infty} - T_{10}) \left(\frac{r + Z_{10}}{r + Z} \right)^2 \cdot \exp(-\lambda \xi) \quad (4b)$$

An evaluation of equation 4a for $Z = 200 \text{ km}$, the top of the region of interest for Task Group II, yields $T = 854.5591 \text{ K}$. Because the model of Task Group III, extending up to 1000 km , uses the same temperature-height profile as that defined by equation 4a, that equation also designates T within that extended height region. For Z equal to 450 km , 500 km , and 1000 km , these temperatures are 998.2247 K , 999.2356 K , and 999.9997 K , respectively. Figure 2 depicts the temperature-height profile between 100 and 450 km .

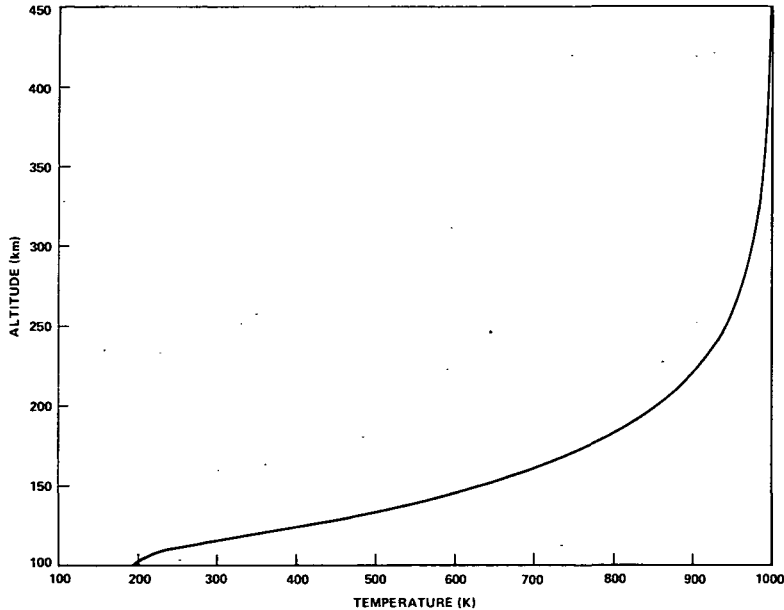


Figure 2. Atmospheric temperature versus altitude, 100 to 450 km.

NUMBER DENSITIES OF INDIVIDUAL SPECIES

General Considerations

The steady-state vertical distribution of a minor gas species of number density n_i (m^{-3}) and molecular weight M_i (kg kmol^{-1}) is governed by the vertical component of the momentum equation for that gas (for example, Colegrove et al., 1965):

$$n_i v_i + D_i \left(\frac{dn_i}{dz} + \frac{n_i(1 + \alpha_i)}{T} \cdot \frac{dT}{dz} + \frac{n_i}{H_i} \right) + K \left(\frac{dn_i}{dz} + \frac{n_i}{T} \cdot \frac{dT}{dz} + \frac{n_i}{H_p} \right) = 0 \quad (5)$$

where

v_i = the flow velocity (m s^{-1}) of the i th species,

D_i = height-dependent, species-dependent, molecular-diffusion coefficient ($\text{m}^2 \text{s}^{-1}$) for the i th minor species diffusing through the major background gas,

α_i = thermal diffusion factor (dimensionless) for the i th species (having a value of -0.4 for He, and zero for O, O_2 , and Ar),

K = height-dependent, eddy-diffusion coefficient ($\text{m}^2 \text{s}^{-1}$).

The quantity H_p is the height-dependent pressure scale height (m) for air

$$H_p \equiv \frac{R^* T}{Mg} \quad (5a)$$

while H_i is the related parameter for a particular gas species, that is,

$$H_i = \frac{R^* T}{M_i g} \quad (5b)$$

where M_i is the molecular weight (kg kmol^{-1}) for the i th species and g is the height-dependent acceleration of gravity (m s^{-2}), which for 45°N , is closely approximated by

$$g = g_0 [r/(r + Z)]^2 \quad (5c)$$

In this expression, $r = 6356.766 \text{ km}$, as previously defined.

Molecular Nitrogen

Equation 5 is used to calculate the distribution of each gas except molecular nitrogen. Since N_2 is the major gas in the lower part of the thermosphere, a different scheme is used to describe its distribution. From the lower boundary altitude of 86 km up to about 100 km, the atmosphere is well mixed as the eddy processes dominate the molecular diffusion, while above about 100 km, molecular diffusion dominates. Also, because $M(\text{N}_2)$, the molecular weight of N_2 , that is, $28.0134 \text{ kg kmol}^{-1}$, is quite close to the mean molecular weight of air, $28.9644 \text{ kg kmol}^{-1}$, the transition from a mixed to a diffusive distribution

of atmospheric gases at heights below which dissociation and diffusion become important, has little effect on the height profile of molecular nitrogen. This allows the use of the following simplified version of equation 5 with the flow velocity (v_i), as well as the eddy-diffusion coefficient (K) and the thermal-diffusion factor (α_i), all set equal to zero. In addition, $n(N_2)$ replaces n_i to represent N_2 number density, and $1/H(N_2)$ replaces $1/H_i$ through which equation 5b is seen to equal $M(N_2) \cdot g/(R^* \cdot T)$:

$$\frac{dn(N_2)}{dZ} + \frac{n(N_2)}{T} \cdot \frac{dT}{dZ} + \frac{n(N_2) \cdot M(N_2) \cdot g}{R^* T} = 0 \quad (6)$$

Since g varies only slowly with height, it is apparent that temperature is the dominant parameter governing the height profile of molecular nitrogen. The solution of equation 6 for $n(N_2)$ yields

$$n(N_2) = n(N_2)_7 \cdot \exp \left[- \int_{z_7}^z \left(\frac{M(N_2) \cdot g}{R^* T} + \frac{1}{T} \cdot \frac{dT}{dZ} \right) dZ \right] \quad (7)$$

where $n(N_2)_7$ is the N_2 number density at level 7 which is 86 km.

The slight change in the height distribution of N_2 caused by the transition from a mixing regime below 100 km to a diffusive-separation regime above this height is accounted for by using the value of the mean molecular weight of air in place of that of $M(N_2)$ at heights below 100 km.

Atomic and Molecular Oxygen, Helium, and Argon

Ideally, equation 5 is solved in conjunction with the equation of continuity (Colegrove et al., 1965; Keneshea and Zimmerman, 1970)

$$\frac{d(n_i v_i)}{dZ} = \zeta(Z) \quad (8)$$

where ζ denotes chemical production and loss terms. For the purpose of the single-profile steady-state model being generated here, however, such a sophisticated and detailed calculation is considered inappropriate. Instead, the flux term, $n_i v_i$, is artificially adjusted† to match the fluxes resulting from the diurnally averaged, time-dependent calculations (Minzner et al., 1976) which account for the effects of photochemical production and loss, as well as for vertical transport, on the vertical distribution of atomic and molecular oxygen. For helium and argon, the flux term represents only the vertical flow.

†Huang, F. T., private communication, 1974.

Equation 5 is integrated directly to obtain the following exponential expression:

$$n_i = n_{i,7} \cdot \frac{T_7}{T} \cdot \exp \left\{ - \int_{Z_7}^Z \left[F(Z) + \left(\frac{v_i}{D_i + K} \right) \right] dZ \right\} \quad (9)$$

where $n_{i,7}$ represents the number density at the lower-level boundary of Model II for any one of the minor species, O, O₂, Ar, and He. The values of $n_{i,7}$, one for each of these minor species as well as one for $n(N_2)_7$, were generated in a manner described in Minzner et al. (1976) and are given in table 2.

The function $F(Z)$ in equation 9 is defined as

$$F(Z) = \frac{D_i}{H_i (D_i + K)} + \left(\frac{\alpha_i D_i}{D_i + K} \right) \frac{1}{T} \cdot \frac{dT}{dZ} + \frac{K}{H_p (D_i + K)} \quad (10)$$

The height profile of the eddy-diffusion coefficient, K , used in the calculation of number density is defined in three segments:

1. For $86 \leq Z < 95$ km,

$$K = 1.2 \times 10^2 \text{ m}^2 \text{ s}^{-1} \quad (11)$$

2. For $95 \leq Z \leq 115$ km,

$$K = 1.2 \times 10^2 \cdot \exp \left(1 - \frac{400}{[400 - (Z - 95)^2]} \right) \quad (12)$$

3. For $Z > 115$ km,

$$K = 0.0 \quad (13)$$

The height-dependent molecular-diffusion coefficients, D_i , for the various species are associated with diffusion through molecular nitrogen, and have the general form

$$D_i = \frac{a_i}{N} \left[\frac{T}{273.15} \right]^{b_i} \quad (14)$$

where N is the total number density at height Z . The value of the coefficients a_i and b_i for various species, O, O₂, Ar, and He (Colegrove et al., 1966) are listed in table 4. With the appropriate values of these coefficients used in equation 14, the resulting value of the function D_i , the molecular-diffusion coefficient, for each of the four indicated species is found to be as plotted in figure 3 for the height interval 86 to 150 km.

Table 4
Values of Species-dependent Coefficients Used in the Expressions
for Molecular-diffusion Coefficients

Gas	$a_i \text{ (m}^{-1} \text{ s}^{-1}\text{)}$	$b_i \text{ (dimensionless)}$
O	6.986×10^{20}	0.750
O ₂	4.863×10^{20}	0.750
Ar	4.487×10^{20}	0.870
He	1.700×10^{21}	0.691

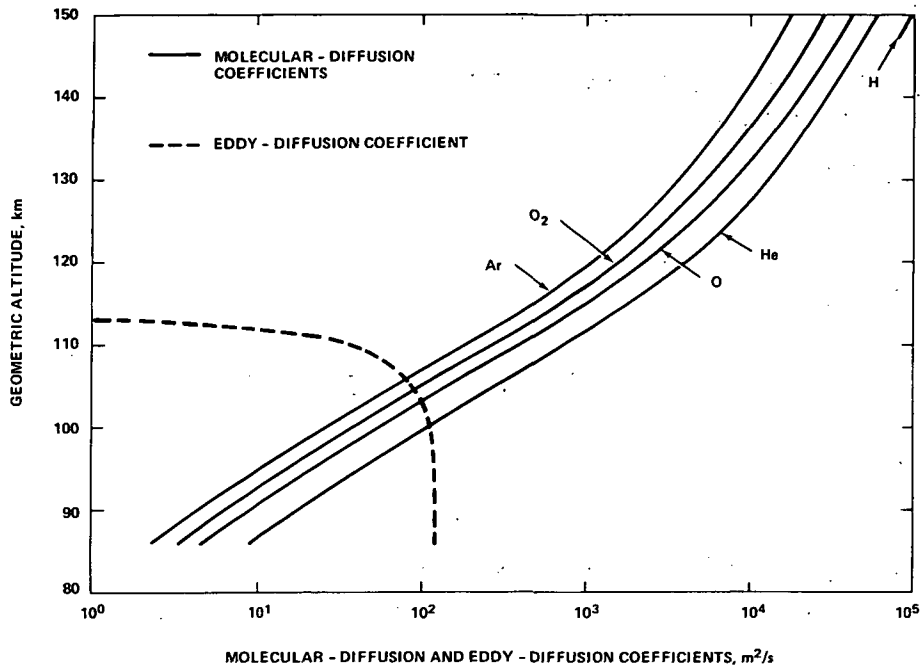


Figure 3. Molecular- and eddy-diffusion coefficients versus altitude, 86 to 150 km.

Figure 3 also shows the shape of the function K and the eddy-diffusion coefficient defined by equations 11, 12, and 13. The significance of K in the height distribution of the number densities of the several species is seen to decrease rapidly above 100 km and to vanish at 115 km. Thus, for $Z > 115$ km, $F(Z)$ defined by equation 10 simplifies to

$$F(Z) = \frac{1}{H_i} + \frac{a_i}{T} \frac{dT}{dZ} \quad (15)$$

and it is obvious that values of D_i no longer enter into the calculation of $F(Z)$, and, hence, not into the calculation of n_i for O, O₂, Ar, and He. Consequently, the height variation of D_i for each of these four species is plotted in figure 3 only for a limited height region, 86 to 150 km. In the calculation of the number density for atomic hydrogen H (see Appendix A), the values of D_i for H are significantly involved to heights far in excess of 150 km. Since $n(H)$ is not calculated for heights below 150 km, however, the graph of D_i for H appears only as a short dash below the 150-km coordinate of figure 3.

The value of the total number density to be used in equation 14 is obtained by sequential calculations, with N_2 providing the value of N for the calculation of the diffusion coefficients for atomic and molecular oxygen, and the sum of the number densities for N_2 , O₂, and O providing the value of N for the calculation of argon and helium diffusion coefficients.

The flux term, $v_i/(D_i + K)$ in equation 9, is represented by the following expression†

$$\frac{v_i}{D_i + K} = Q_i (Z - U_i)^2 \cdot \exp[-W_i (Z - U_i)^3] + q_i (u_i - Z)^2 \cdot \exp[-w_i (u_i - Z)^3] \quad (16)$$

The coefficients q_i , Q_i , u_i , U_i , w_i , and W_i , which are constant for a particular species, are each adjusted such that appropriate densities are obtained at 450 km for O and He, and at 150 km for O, O₂, He, and Ar. The constant, q_i , is zero for all species except atomic oxygen, and is also zero for atomic oxygen above 97 km; the extra term for atomic oxygen is needed below 97 km to generate a maximum in the density-height profile at the selected height of 97 km. This maximum results from the increased loss of atomic oxygen by recombination at lower altitudes. The flux terms for O and O₂ are based on, and lead (qualitatively) to, the same results as those derived from the much more detailed calculations by Colegrove et al. (1965) and Keneshea and Zimmerman (1970). Table 5 lists the values of the coefficients q_i , Q_i , u_i , U_i , w_i , and W_i , for each of the four minor species being considered, while the lower boundary-condition number-density values required in equation 5 are included in table 2.

Table 5
Values of Species-dependent Coefficients Applicable to the Empirical, Integrable
Equation 16 Representing the Flux Term $v_i/(D_i + K)$

Gas	q_i (km ⁻³)	Q_i (km ⁻³)	u_i (km)	U_i (km)	w_i (km ⁻³)	W_i (km ⁻³)
O	-3.416248×10^{-3}	-5.809644×10^{-4}	97.0	56.90311	5.008765×10^{-4}	2.706246×10^{-5}
O ₂	0	1.366312×10^{-4}	—	86.000	—	8.333333×10^{-5}
Ar	0	9.434079×10^{-5}	—	86.000	—	8.333333×10^{-5}
He	0	-2.457369×10^{-4}	—	86.000	—	6.666667×10^{-4}

†Huang, F. T., private communication, 1974.

Number densities versus height for each of N_2 , O, O_2 , Ar, and He were computed using equations 6 through 16. Graphs of these data for each species were prepared for each of two height regions, 86 to 150 km and 100 to 450 km. These graphs are presented in figures 4 through 13.

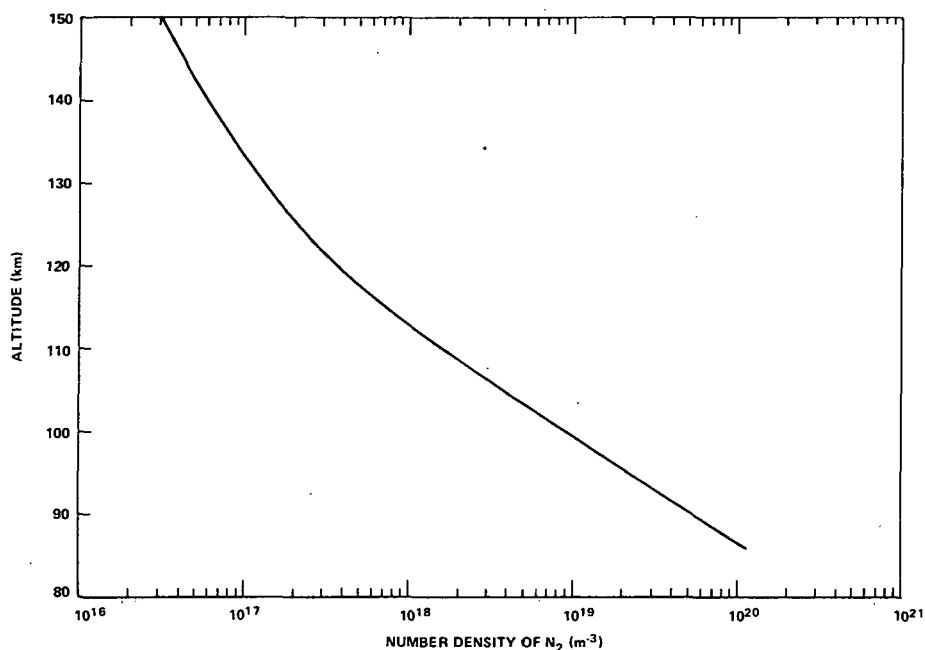


Figure 4. Number density of N_2 versus altitude, 86 to 150 km.

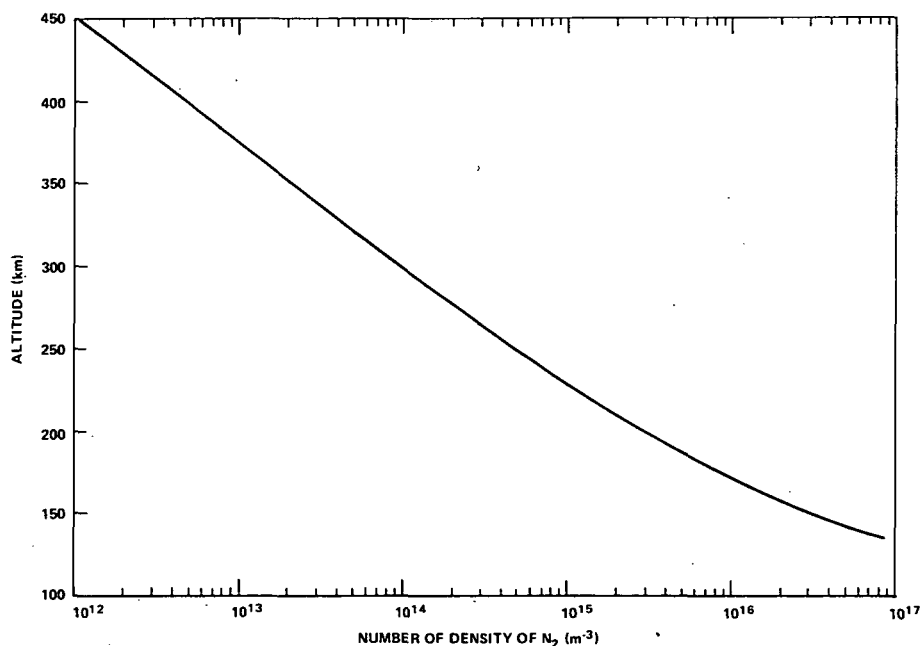


Figure 5. Number density of N_2 versus altitude, 100 to 450 km.

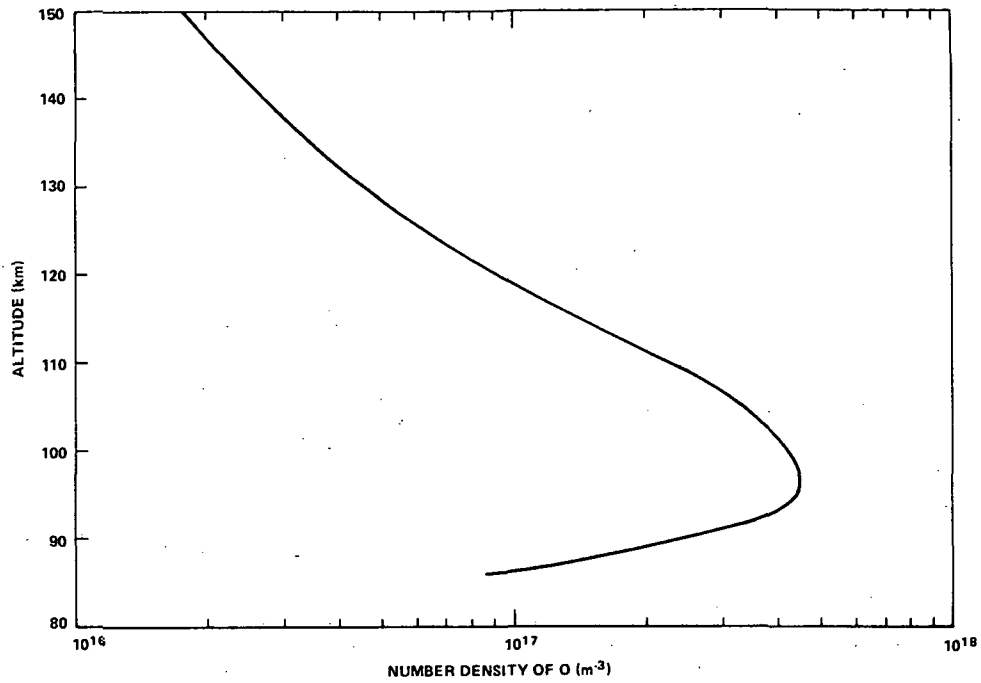


Figure 6. Number density of O versus altitude, 86 to 150 km.

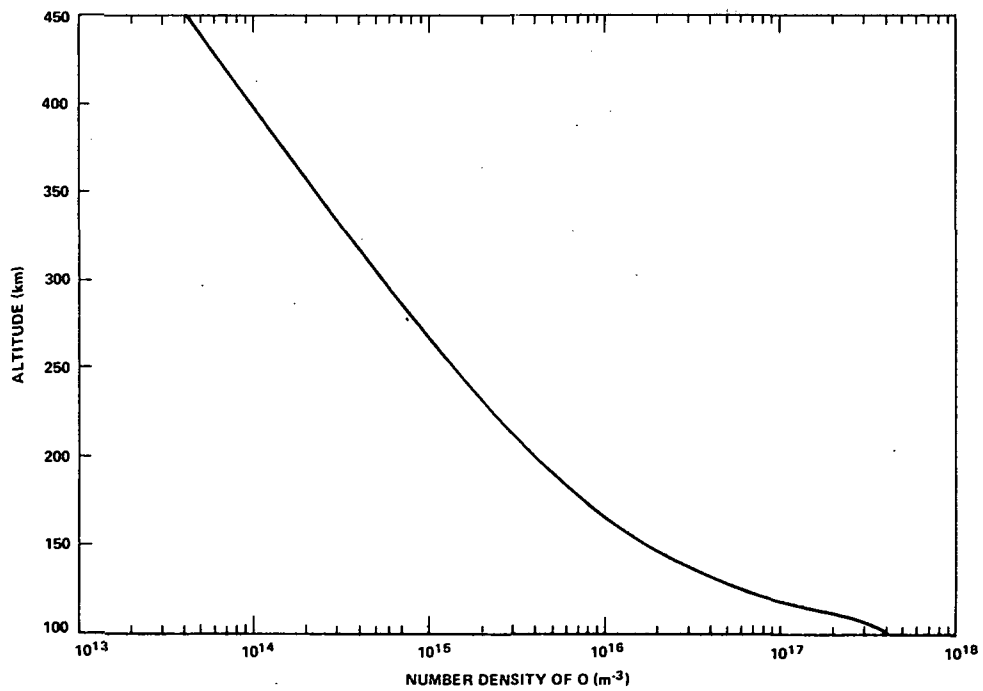


Figure 7. Number density of O versus altitude, 100 to 450 km.

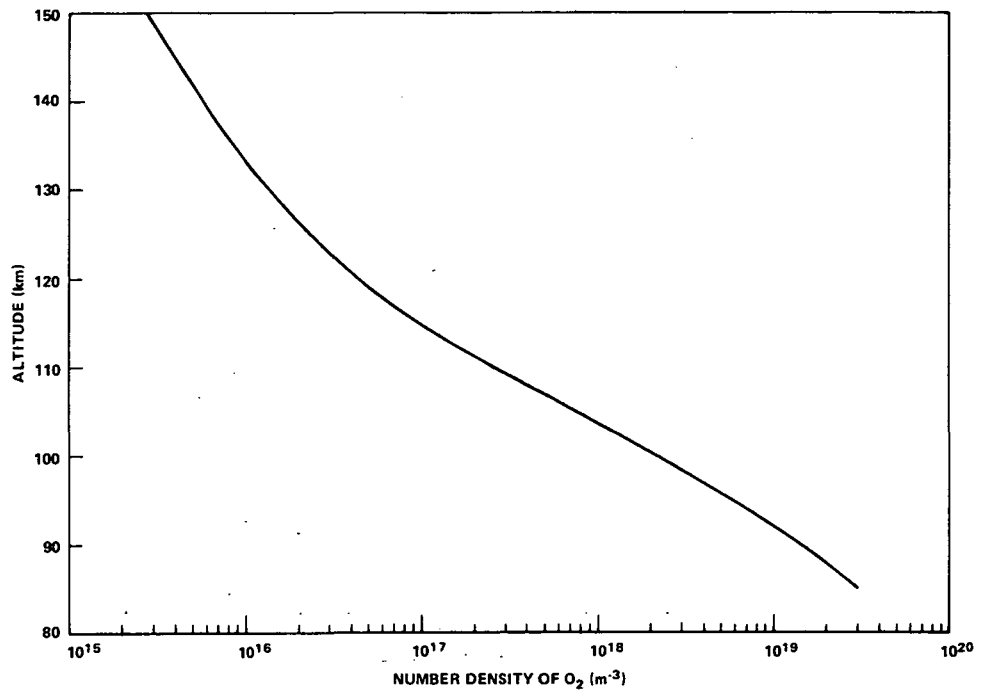


Figure 8. Number density of O_2 versus altitude, 86 to 150 km.

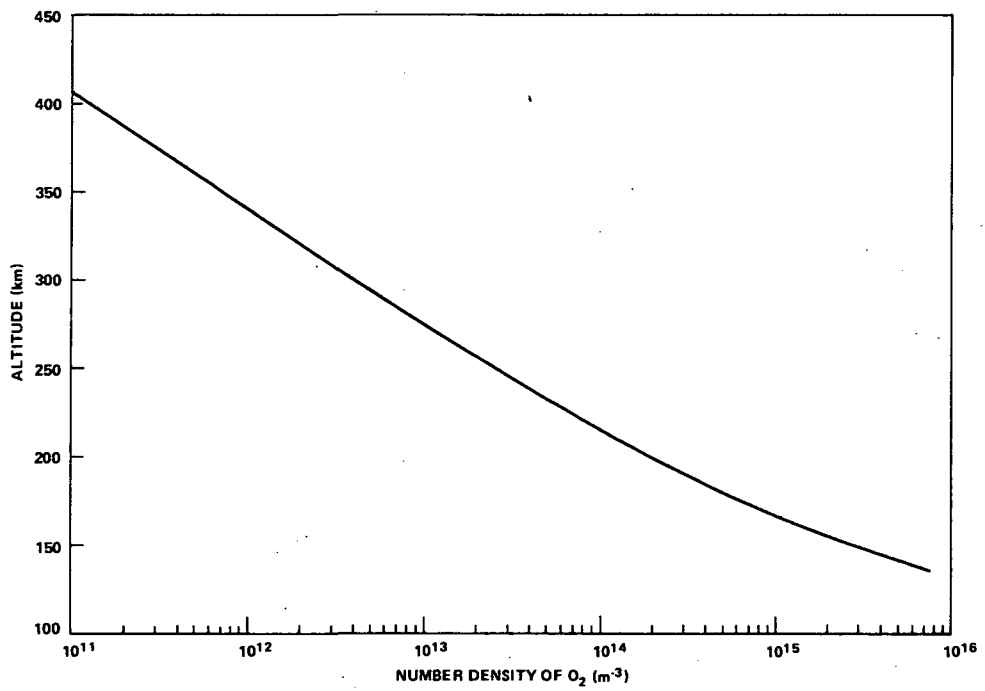


Figure 9. Number density of O_2 versus altitude, 100 to 450 km.

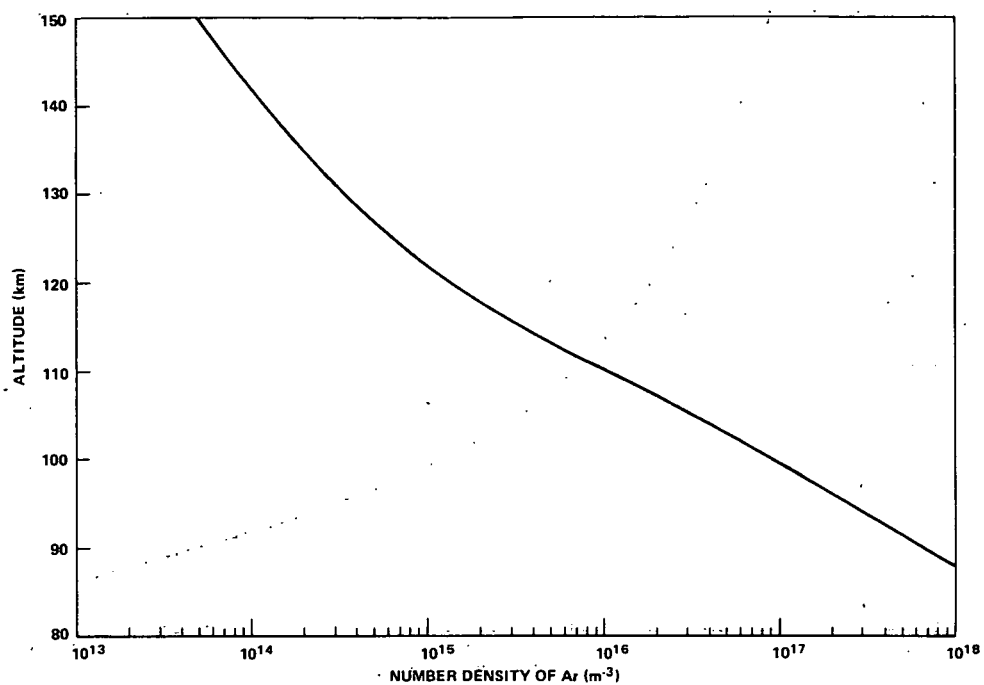


Figure 10: Number density of Ar versus altitude, 86 to 150 km.

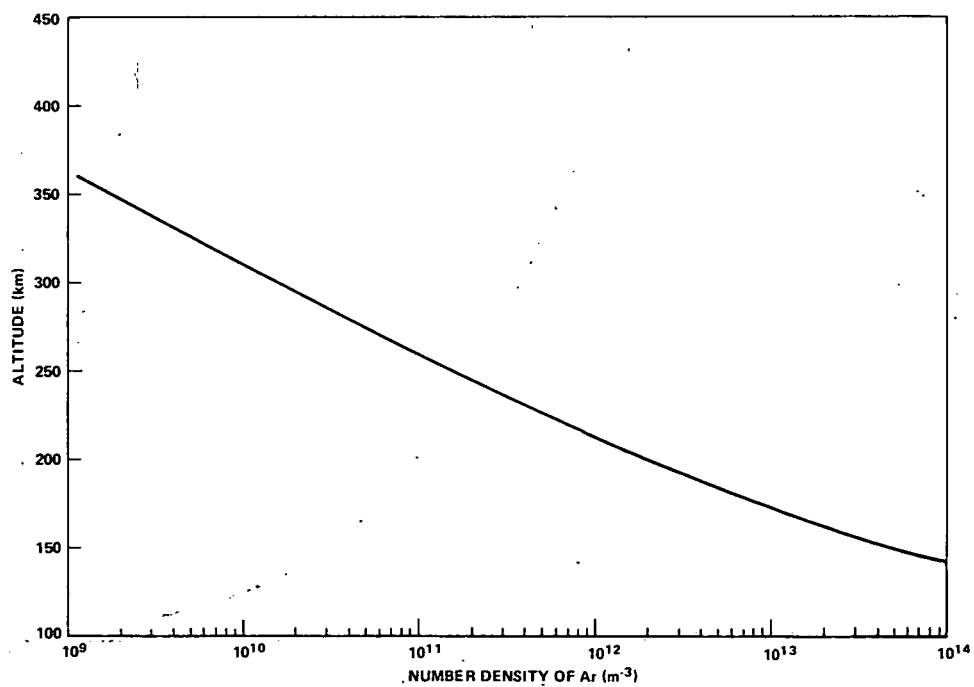


Figure 11: Number density of Ar versus altitude, 100 to 450 km.

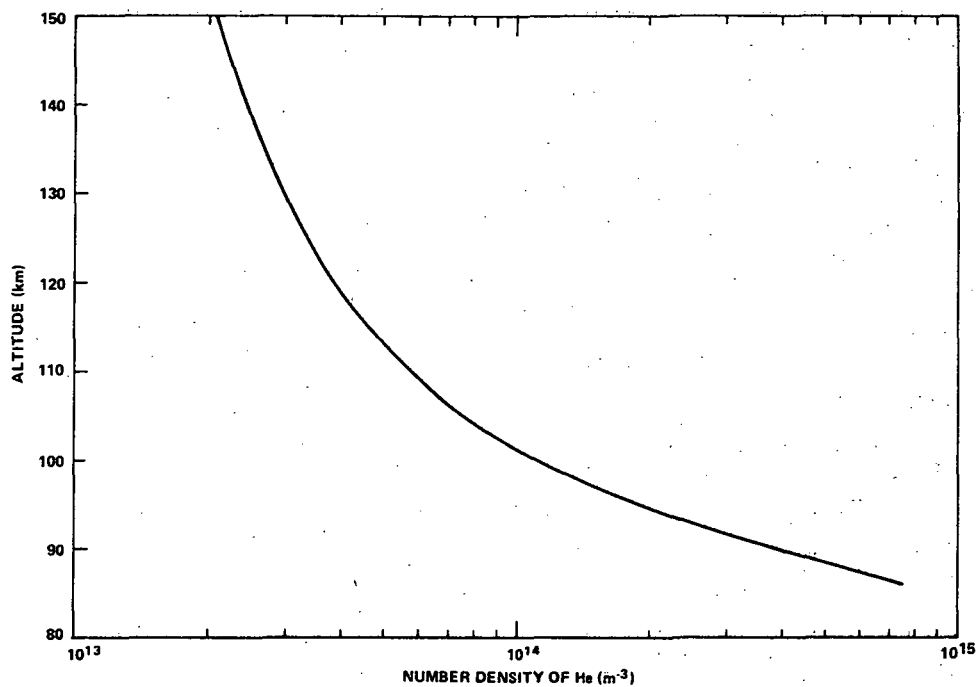


Figure 12. Number density of He versus altitude, 86 to 150 km.

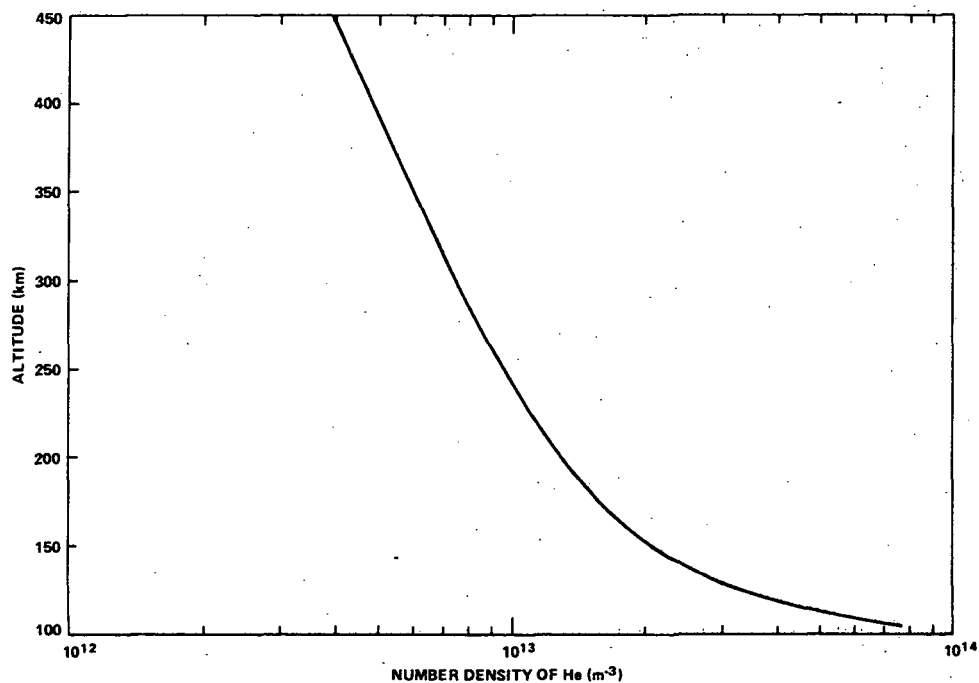


Figure 13. Number density of He versus altitude, 100 to 450 km.

Atomic Hydrogen

A set of calculated atomic-hydrogen number densities versus height, originally prepared for Task Group III for heights from 150 to 1000 km, were supplied to Task Group II by Forbes, in punch-card form. The method for calculating these values is given in Appendix A. The atomic-hydrogen height profile generated from the supplied values is shown in figure 14 for heights between 150 and 450 km, and has a value of 3.7541×10^{11} at 150 km. This value represents about one part in 100,000 of the N_2 number density (3.1221×10^{16}) at that height. Consequently, at heights below 150 km, for which height region no atomic-hydrogen number densities were given, these values would provide a negligible contribution to the four-significant-figure value of total number density N , and need not be considered in its calculation at these heights.

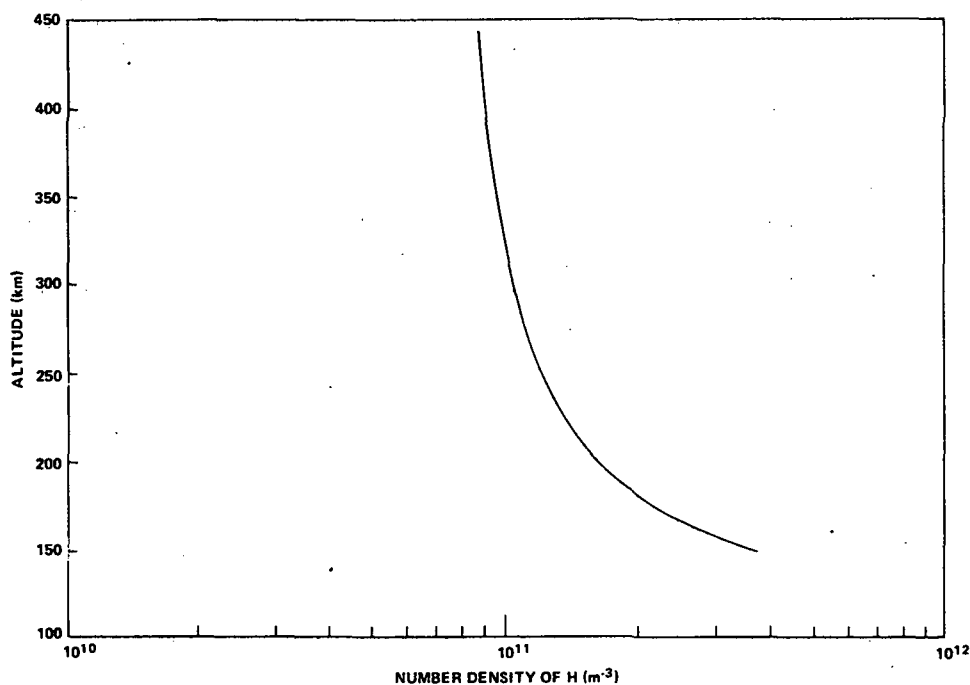


Figure 14. Number density of H versus altitude, 150 to 450 km.

TOTAL NUMBER DENSITY

The total number density, N (m^{-3}), at each of the tabulated heights from 86 through 149 km, consists of the sum of the number densities, n_i , of the individual species N_2 , O , O_2 , Ar , and He , at each of the corresponding heights, that is,

$$N = \sum n_i \quad (17)$$

For heights of 150 km and above, N includes n_i for atomic hydrogen. Graphs of total number density versus height in each of two height ranges, 86 to 150 km and 100 to 450 km, are presented in figures 15 and 16, respectively.

TOTAL MASS DENSITY

The total mass density, ρ (kg m^{-3}), for heights from 86 through 450 km is determined from the number density, n_i , of the five or six recognized constituents at each height in that height regime, using the relationship

$$\rho = \frac{\sum (n_i M_i)}{N_A} \quad (18)$$

where M_i is the molecular weight of the i th species as listed in the *U.S. Standard Atmosphere, 1962* (COESA, 1962), that is,

- 28.0134 kg kmol^{-1} for N_2 ,
- 31.9988 kg kmol^{-1} for O_2 ,
- 15.9994 kg kmol^{-1} for O ,
- 39.948 kg kmol^{-1} for Ar ,
- 4.0026 kg kmol^{-1} for He ,
- 1.00797 kg kmol^{-1} for H ,
- N_A is Avogadro's constant $6.022169 \times 10^{26} \text{ kmol}^{-1}$.

Graphs of ρ versus height in two height ranges, 86 to 150 km and 100 to 400 km, are given in figures 17 and 18, respectively.

MEAN MOLECULAR WEIGHT

The mean molecular weight, M (kg kmol^{-1}), at height Z from 86 to 450 km is determined from the number densities, n_i , of the five or six recognized species constituting the atmosphere in this height regime, using the relationship

$$M = \frac{\sum (n_i M_i)}{\sum n_i} \quad (19)$$

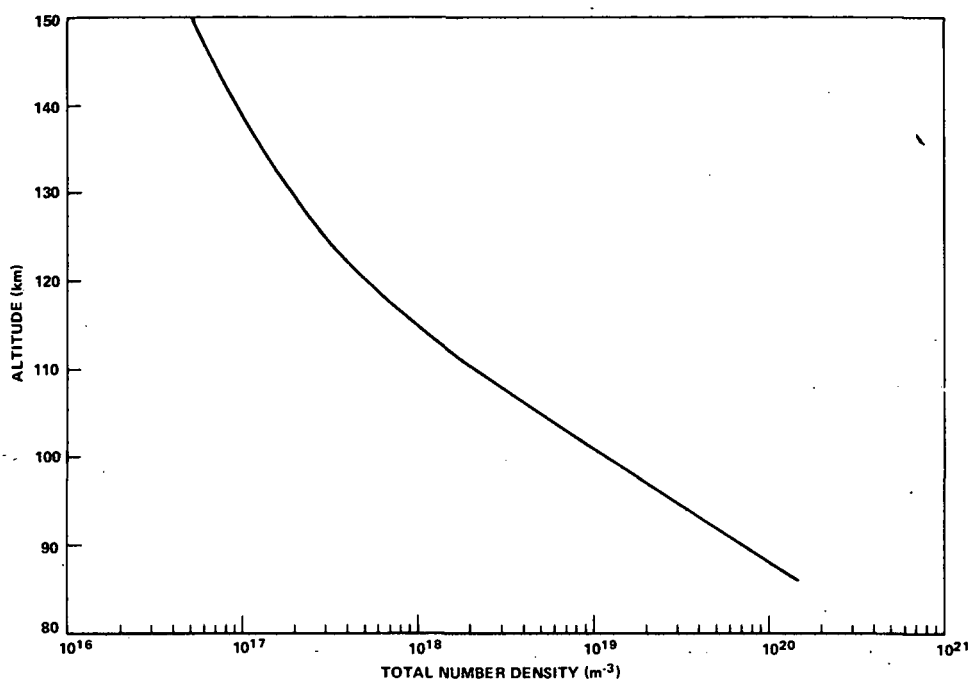


Figure 15. Total number density versus altitude, 86 to 150 km.

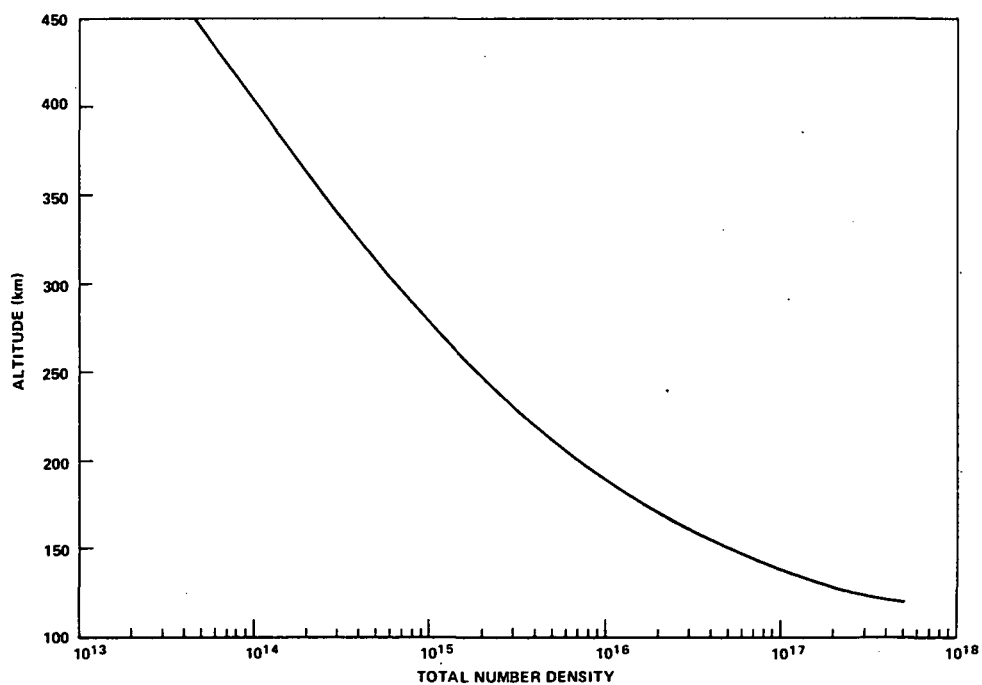


Figure 16. Total number density versus altitude, 100 to 450 km.

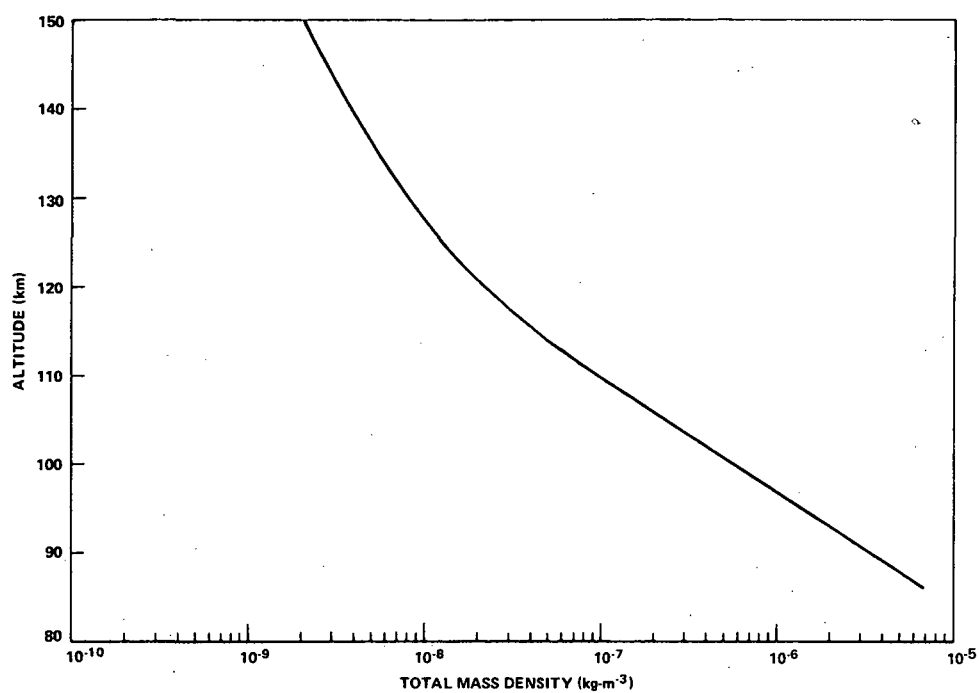


Figure 17. Total mass density versus altitude, 86 to 150 km.

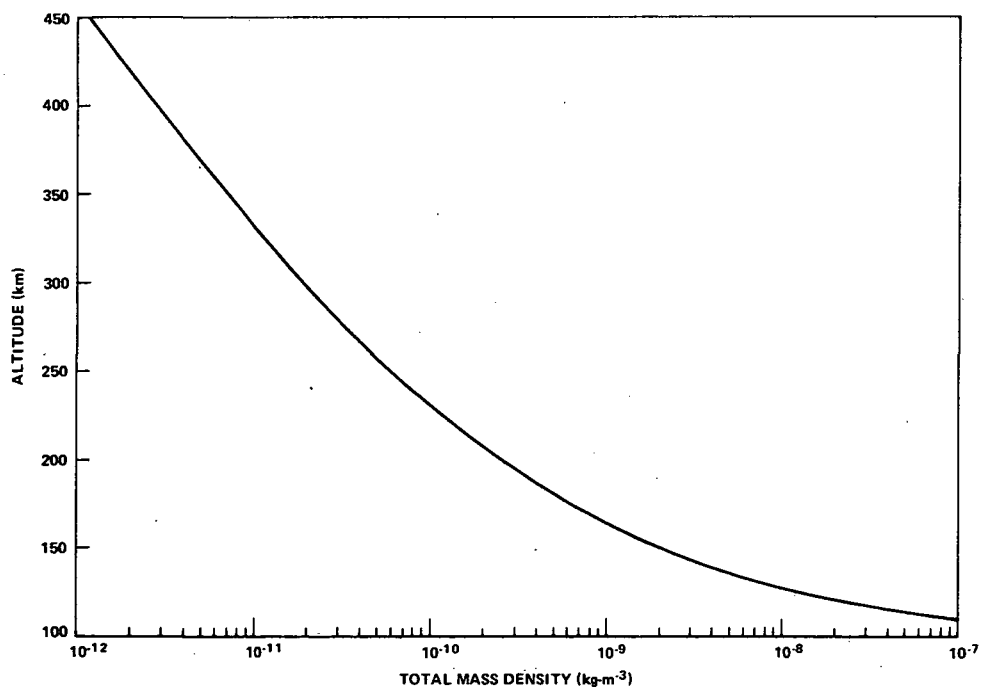


Figure 18. Total mass density versus altitude, 100 to 450 km.

Graphs of mean molecular weight versus height in the ranges of 86 to 150 km and 100 to 450 km are given in figures 19 and 20, respectively.

TOTAL PRESSURE

The total pressure, P , in N m^{-2} or equivalently in pascals (Pa), at heights of 86 km and above is computed from the mass density, ρ , the kinetic temperature, T , and the mean molecular weight, M , using the relationship

$$P = \frac{\rho R^* T}{M} \quad (20)$$

where R^* is the universal gas constant, $8.31432 \times 10^3 \text{ (J K}^{-1} \text{ kmol}^{-1})$. Because ρ is a function of the value of n_i for each of the six constituents, P is also dependent primarily upon these quantities. To convert from N m^{-2} or Pa to millibars (mbar), the customary meteorological unit for pressure, one must divide the results of equation 20 by 100. Graphs of total pressure in millibars as a function of height are given in figures 21 and 22.

SCALE HEIGHT

The total-pressure scale height, H_p (m), more commonly called the scale height, is related to mean molecular weight and temperature as well as to the acceleration of gravity, g , in the relationship

$$H_p = \frac{R^* T}{gM} \quad (21)$$

This quantity was previously defined in relation to the calculation of diffusion coefficients. Graphs of H_p as a function of Z , for $86 \leq Z \leq 150$ and for $100 \leq Z \leq 450$ km are given in figures 23 and 24, respectively.

MEAN PARTICLE SPEED

The mean particle speed, V (m s^{-1}), is related to the ratio R^*T/M , and as such is equal to the product of a constant $(8/\pi)^{1/2} \times (g \cdot H_p)^{1/2}$. The computational expression for mean particle speed is

$$V = \left[\frac{8R^*T}{\pi M} \right]^{1/2} \quad (22)$$

Graphs of V as a function of Z in each of two different height ranges are given in figures 25 and 26.

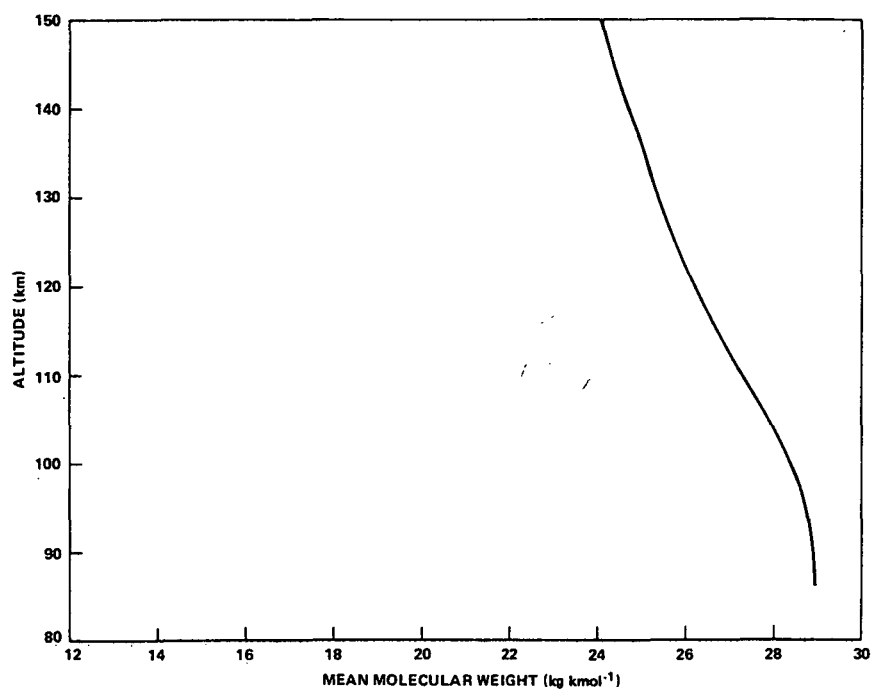


Figure 19. Mean molecular weight versus altitude, 86 to 150 km.

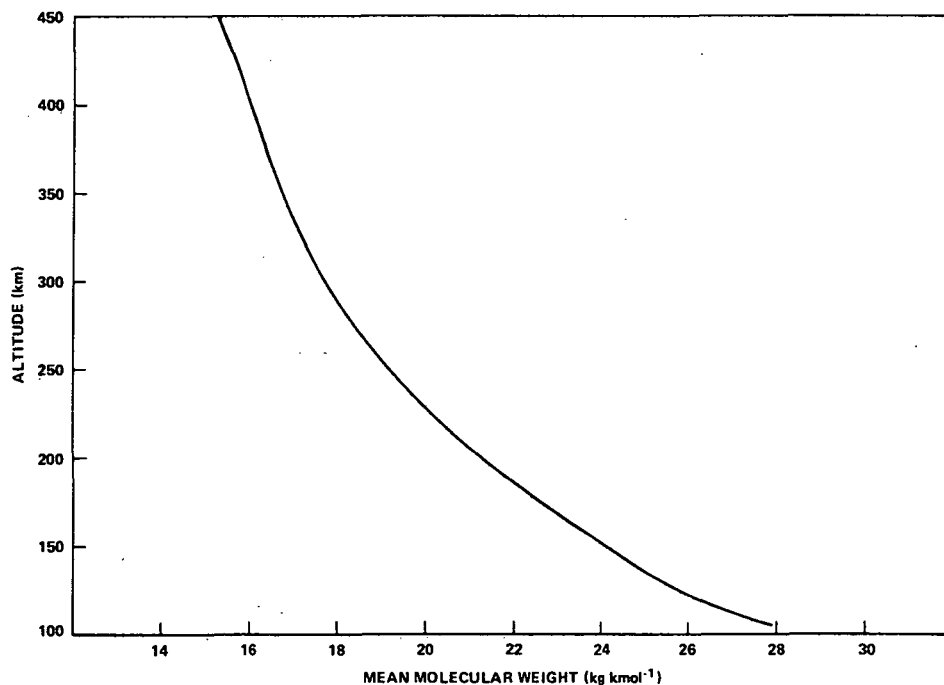


Figure 20. Mean molecular weight versus altitude, 100 to 450 km.

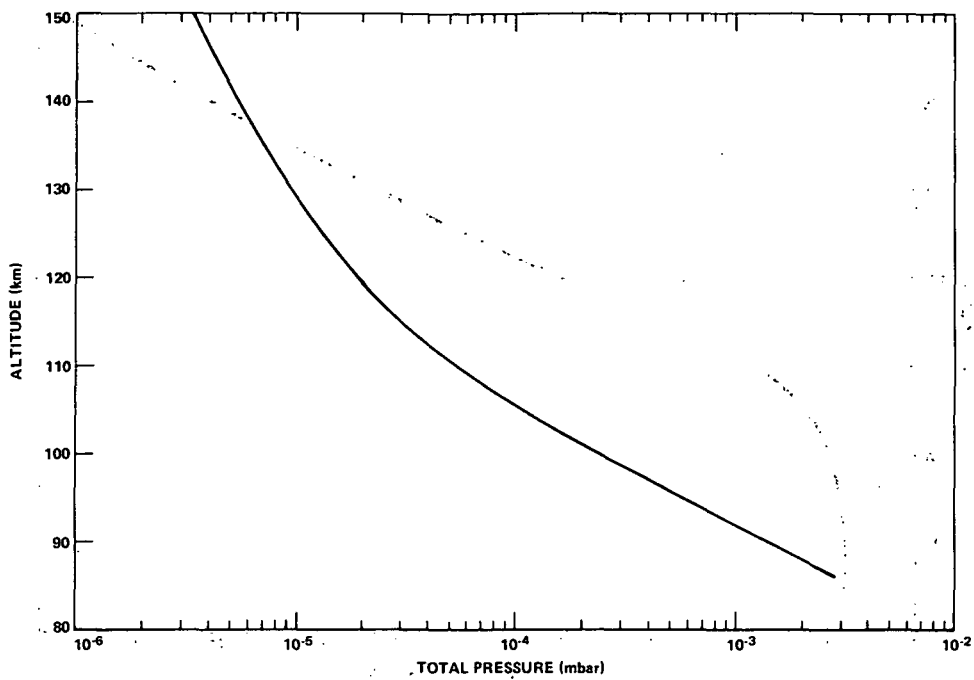


Figure 21. Total pressure versus altitude, 86 to 150 km.

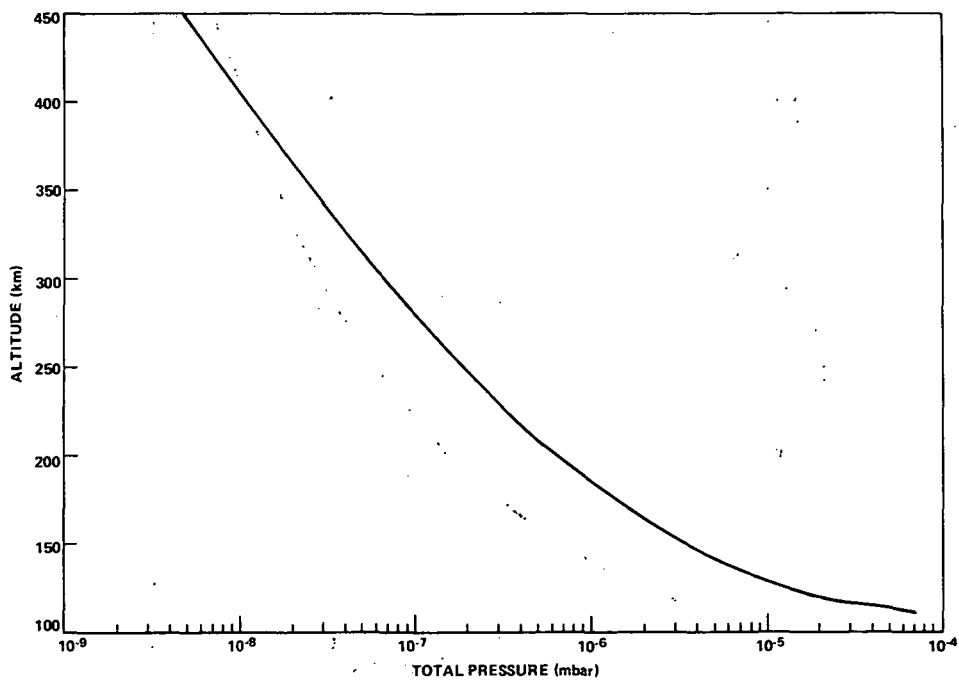


Figure 22. Total pressure versus altitude, 100 to 450 km.

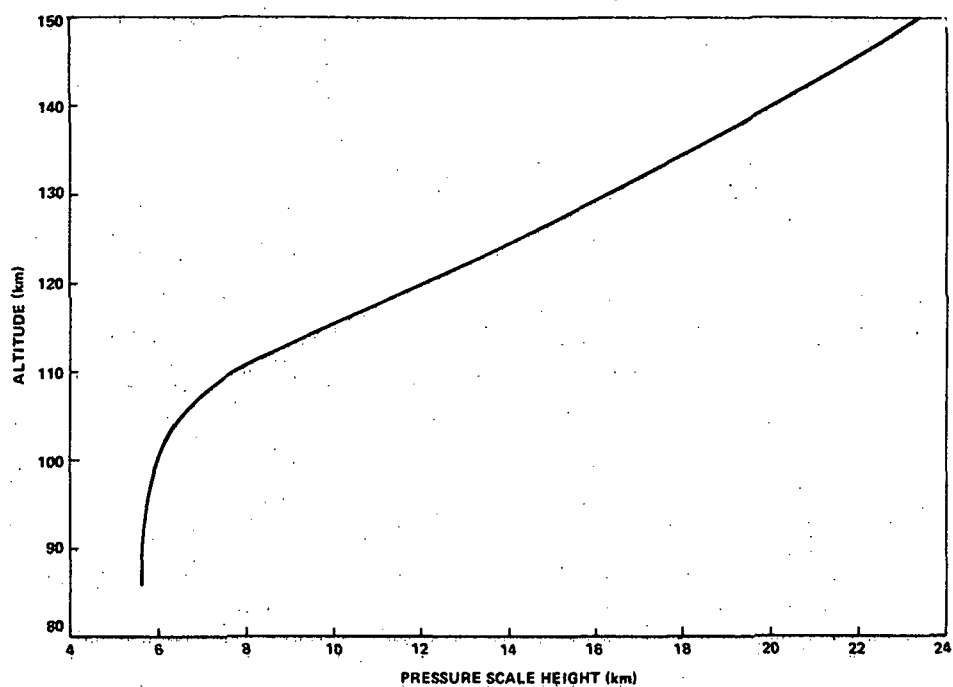


Figure 23. Pressure scale height versus altitude, 86 to 150 km.

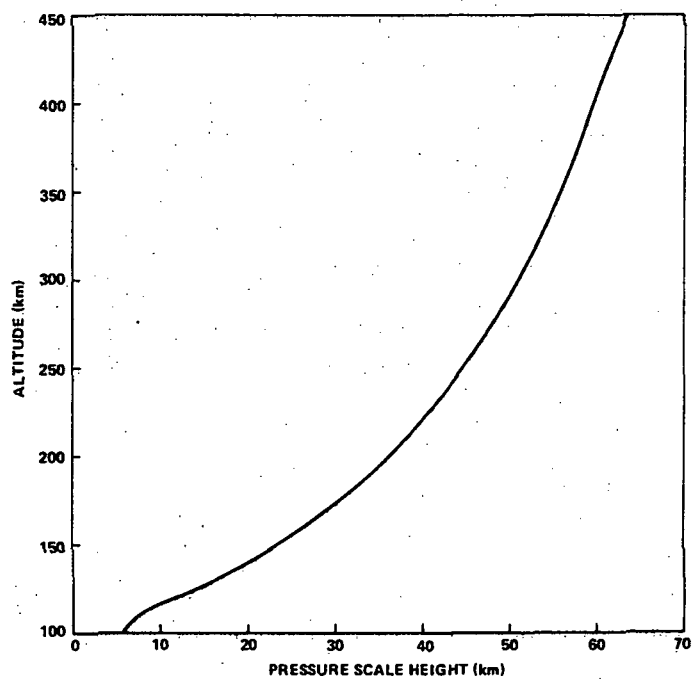


Figure 24. Pressure scale height versus altitude, 100 to 450 km.

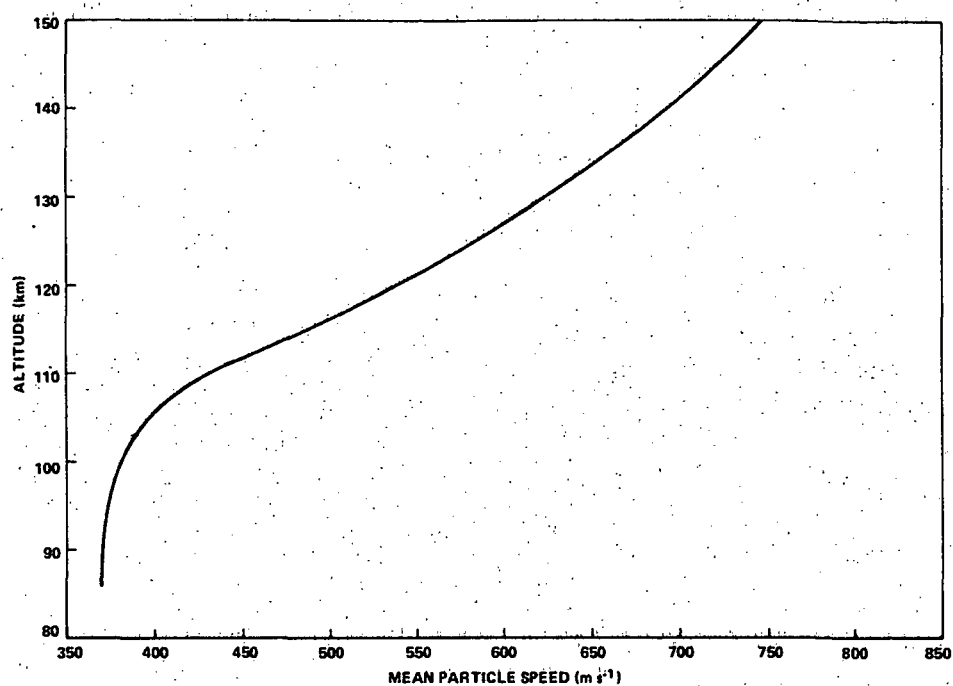


Figure 25. Mean particle speed versus altitude, 86 to 150 km.

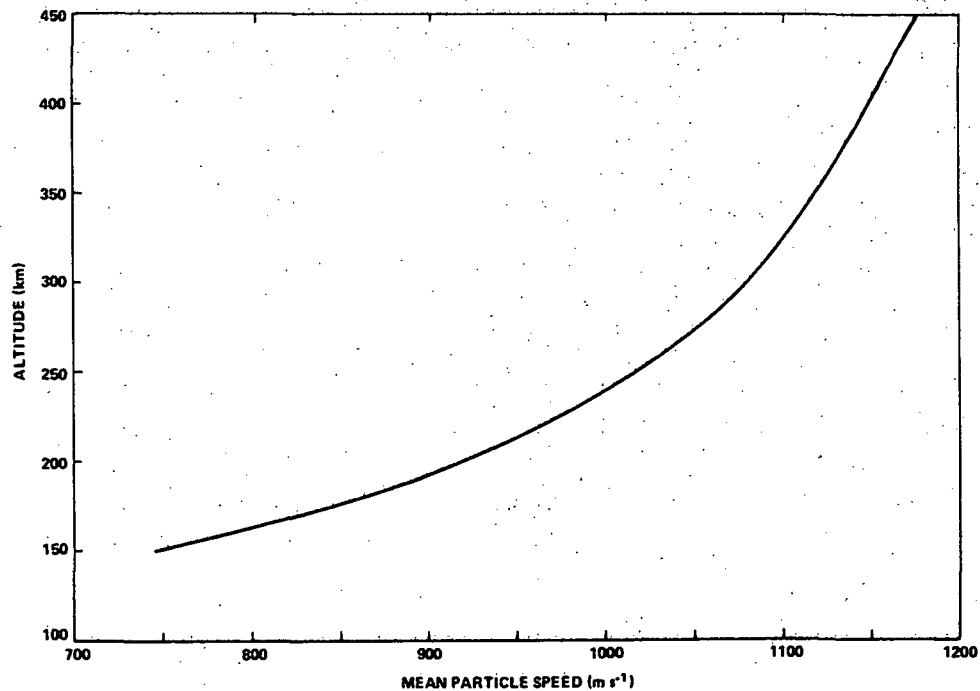


Figure 26. Mean particle speed versus altitude, 100 to 450 km.

MEAN COLLISION FREQUENCY

The mean collision frequency, ν (s^{-1}), between the hypothetical mean neutral particles is a function of P , T , and M , as well as of an effective mean collision diameter, σ . The expression used for computing this property is

$$\nu = 4\sigma^2 N_A P \left(\frac{\pi}{MR^*T} \right)^{1/2} \quad (23)$$

where

π = the ratio of the circumference to the diameter of a circle, that is, 3.14159 (dimensionless)

σ = the effective collision diameter, 3.65×10^{-10} m.

Graphs of this quantity are presented in the height range from 86 to 150 km and from 100 to 450 km in figures 27 and 28, respectively.

MEAN FREE PATH

The mean free path, L (m), which implies a knowledge of σ , is equal to the ratio of mean particle speed to mean collision frequency,

$$L = \frac{V}{\nu} \quad (24)$$

This quantity represents the average travel distance between collisions of the hypothetical mean neutral particles in any direction along which the total change of number density is small over the mean free path length. Thus, at heights below 160 km, where L is less than 10 m, the concept is applicable in all directions. At heights of 500 km and above, however, where L is of the order of 100 km or greater, the concept of mean free path, at best, becomes blurred, and it must be used guardedly. Graphs of L versus Z in the height range of 86 to 150 km and 100 to 450 km are given in figures 29 and 30, respectively.

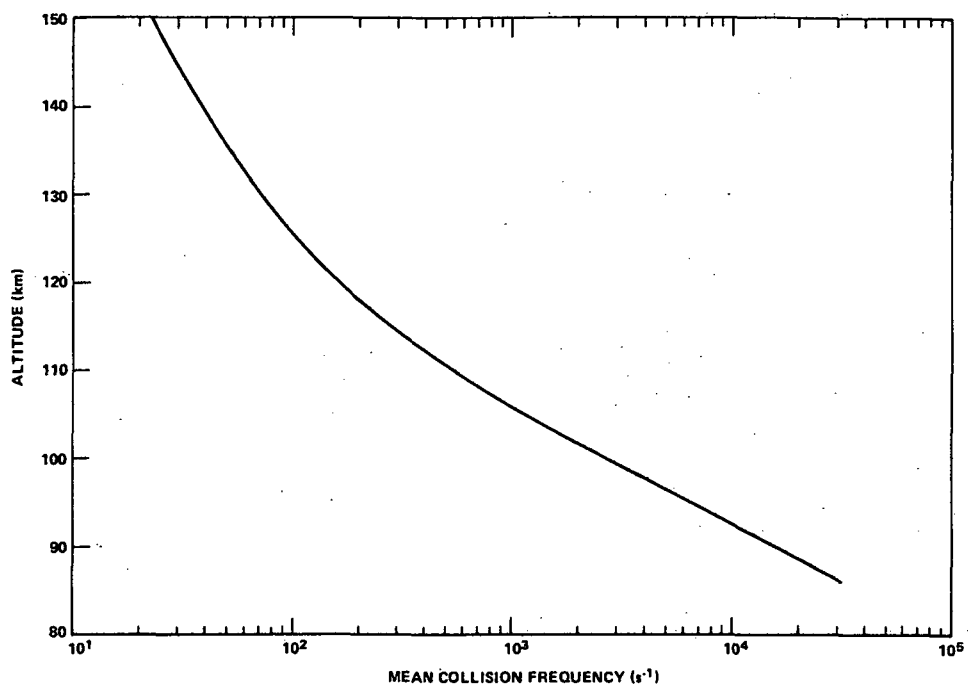


Figure 27. Mean collision frequency versus altitude, 86 to 150 km.

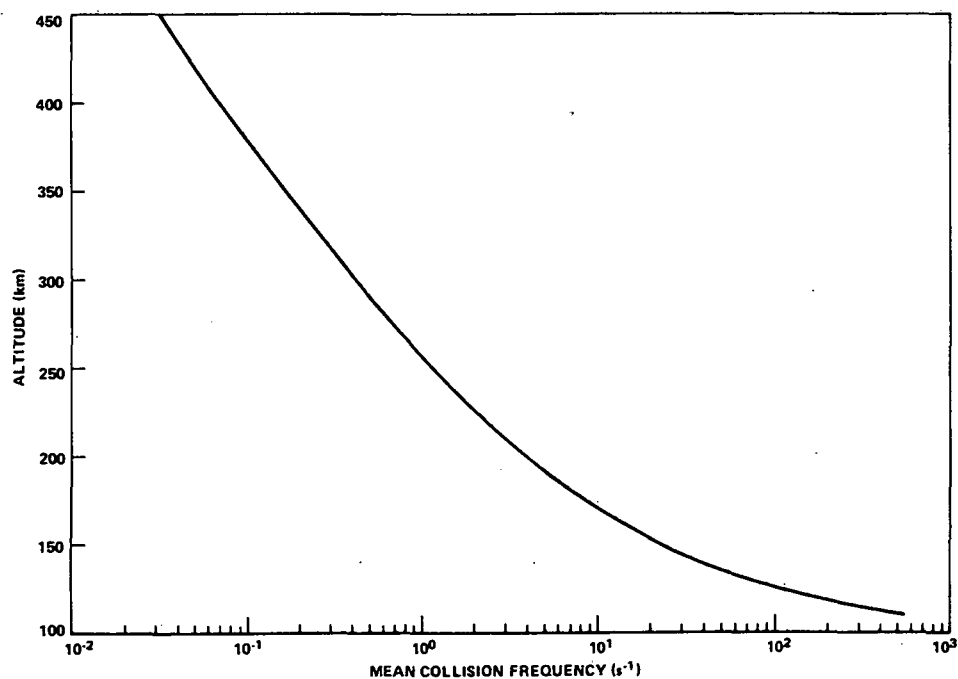


Figure 28. Mean collision frequency versus altitude, 100 to 450 km.

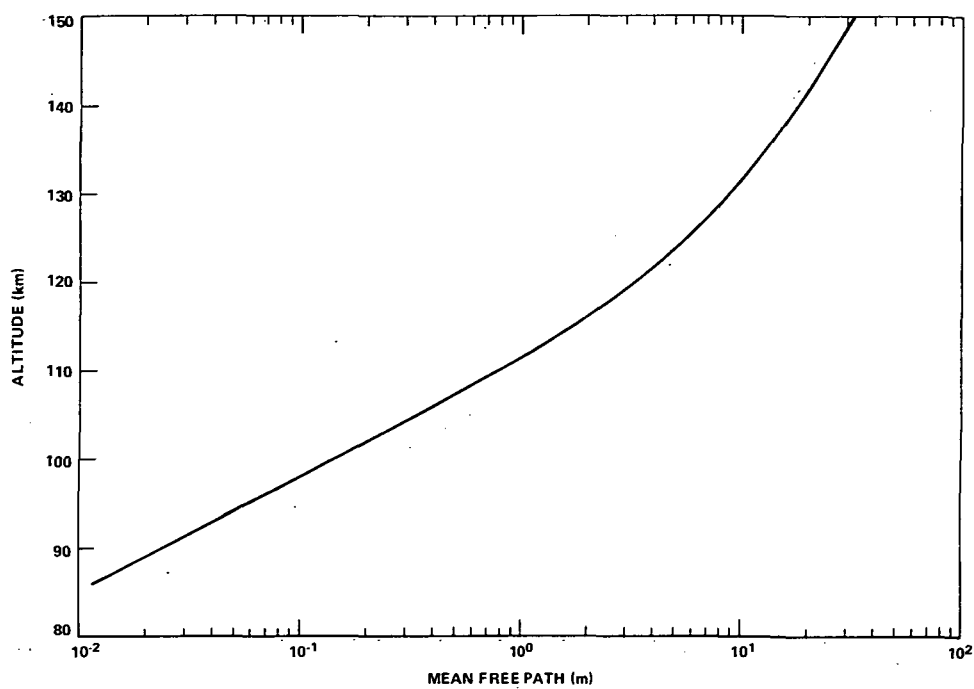


Figure 29. Mean free path versus altitude, 86 to 150 km.

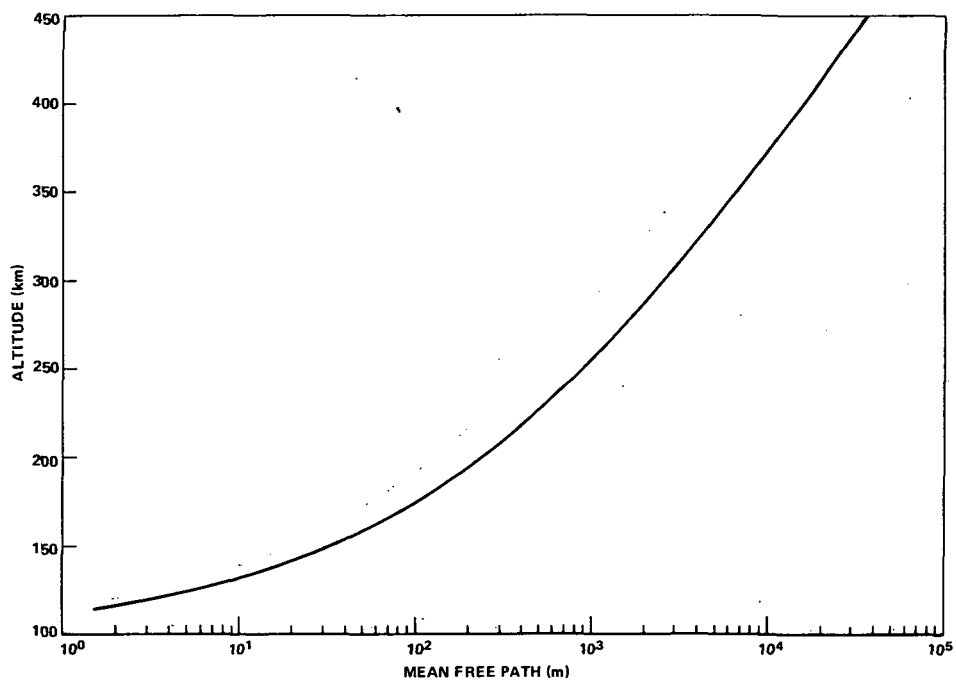


Figure 30. Mean free path versus altitude, 100 to 450 km.

THE MODEL COMPARED WITH OBSERVATIONS

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HEIGHT PROFILES OF TEMPERATURE AND N_2 NUMBER DENSITY

As previously noted, the N_2 number density at any altitude is sensitive primarily to the temperature at lower altitudes. This fact has serious implications when there are a number of data sets to be matched, as in the development of Model II, wherein

- The lower boundary conditions of temperature and density (at 86 km) are determined by Model I.
- The 150-km density is specified from observational data.
- There are some temperature data for the region between 86 and 150 km.

The temperature data come mainly from recent pitot-tube measurements† and from incoherent scatter data.‡ These two data sets are quite consistent in one particular feature: the mean temperature profile between about 105 and 125 km appears to have a constant gradient in each data set, with an average value of about 15 K km^{-1} for the backscatter data, and about 18 K km^{-1} for the pitot-tube data. The adopted version of Model II also exhibits a constant gradient in this height region, but it is 12 K km^{-1} , only two thirds of the larger of these two measured values. Attempts to incorporate higher gradients lead to unacceptably high values for N_2 densities above 150 km.

†Theon, J. and J. Horvath, private communication, 1973.

‡Wand, R., private communication, 1973.

The lower boundary parameters and the N_2 density at 150 km reflect the results of many measurements in which there is a high degree of confidence, so it is unlikely that these data have serious error. It is not clear whether the recent measurements of the temperature-height profile between 110 and 120 km suggest a gradient which is too large, or whether the three inputs are basically inconsistent in that they are not true averages over the same set of conditions.

DYNAMIC CHARACTERISTICS

As noted earlier, most of the properties being modeled are time-dependent by nature, and any steady-state description has to be used advisedly. Examples are the diurnal photochemical variations in atomic and molecular oxygen densities, and the longer term, dynamically induced variations in helium and argon densities. The model described by the preceding equations includes the provision for representing deviations from diffusive equilibrium profiles in the middle thermosphere, whose deviations are becoming more and more accepted as being physically real. Basic considerations for computing a time-dependent model are discussed elsewhere (Minzner et al., 1976).

COMPOSITION

Atmospheric densities in the altitude range of 100 to 200 km, computed from composition measurements made with rocket-borne mass spectrometers, have always been lower than values inferred by downward extrapolation of drag measurements on satellites having orbital altitudes greater than 200 km. While it has been recognized that there might be some error in the drag coefficients upon which the drag-derived densities depended, the general feeling has been that the composition measurements were in error. In particular, because of the highly reactive nature of atomic oxygen, it has been assumed that this constituent was largely lost in mass-spectrometer ion sources, and hence grossly underestimated. Early mass-spectrometric values such as those of Meadows and Townsend (1960) or Pokhunkov (1960) were extremely low, undoubtedly owing to the loss of atomic oxygen on the extensive surfaces of their instruments. With the advent of open source instruments, such as those of Schaefer (1963) and of Nier et al. (1964), much higher values were obtained. Even so, it was recognized that the losses might still be considerable.

Hall et al. (1965, 1967), using EUV extinction measurements made with rocket-borne ultraviolet (UV) spectrometers, found atomic-oxygen abundances in the 150- to 200-km altitude range to be considerably above those reported from rocket-borne mass-spectrometer measurements. Results extrapolated downward from Orbiting Solar Observatory (OSO)-III measurements (Hinteregger and Hall, 1969) gave similar results. The absolute numbers given are in some doubt, however, in view of the uncertainty in the absorption cross section employed for atomic oxygen (Moe, 1970).

The situation at 150 km, as of 1970, was summarized by von Zahn (1970), making use of the relevant published composition values, mass-spectroscopic as well as UV extinctions, available at the time. After evaluating the data, he concluded that the most consistent agreement between densities determined from drag acceleration and those determined from mass spectroscopy was obtained if one assumed that drag-determined densities were high by 10 percent and the values of atomic oxygen found by mass spectroscopy were low by an appreciable factor, perhaps as much as 4. Accordingly, he recommended particle densities at 150 km as follows:

$$n(\text{N}_2) = 2.6 \times 10^{16} \text{ m}^{-3}$$

$$n(\text{O}_2) = 2.5 \times 10^{15} \text{ m}^{-3}$$

$$n(\text{Ar}) = 5 \times 10^{13} \text{ m}^{-3}$$

$$n(\text{O}) = 2.3 \times 10^{16} \text{ m}^{-3}$$

and mass density $\rho = 1.96 \times 10^{-9} \text{ kg/m}$, where his $n(\text{O})$ value was an upward adjustment from available observed values. His drastic increase in the amount of atomic oxygen over observed values seemed justified in part by later measurements made at 120 km with a helium-cooled, rocket-borne mass spectrometer which gave an appreciably higher value for the concentration ratio of O to O_2 at 120 km than had ever been reported in the literature (Offerman and von Zahn, 1971).

Nier (1972), on the other hand, pointed out that since atomic oxygen is a major constituent of the atmosphere in the neighborhood of 150 km, any arbitrary increase, such as by a factor of substantially more than two in its measured abundance relative to other constituents, would destroy the excellent agreement between mass-density scale heights computed from mass spectrometer composition measurements, and those found from drag measurements on low altitude satellites such as OV1-15 (Champion et al., 1970a) and OV1-16 (Champion et al., 1970b). He subsequently reinforced his argument through laboratory experiments (Nier et al., 1972; Lake and Nier, 1973), in which it was shown that it was not likely that atomic-oxygen densities measured with instruments such as he and his colleagues previously used in rocket flights were low by a factor of more than two.

More recently, Taeusch and Carignan (1972), in an extrapolation of OGO-6 composition and drag-determined densities down to 150 km, concluded that the 150-km atomic-oxygen value given by von Zahn (1970), and employed by Jacchia (1971) in his 1971 model, was too high. They prefer a number about 20 percent lower, but still considerably above the average value found with rocket-borne mass spectrometers. Their $n(\text{N}_2)$ and $n(\text{O}_2)$ values at 150 km, on the other hand, are about 25 percent higher than values generally found with rocket-borne mass spectrometers.

Moe (1973) completed a comprehensive study of drag measurements with satellites, as well as of published values of atmospheric composition by all methods, correcting drag measurements for effects due to accommodation coefficients, and composition measurements for possible errors in instruments due to surface effects. Moe's $n(\text{N}_2)$, $n(\text{O}_2)$, and $n(\text{Ar})$ values at 150 km agree closely with those given by von Zahn (1970), which are essentially the

abundances found by rocket-borne mass spectrometers. His $n(\text{O})$ value, however, is about 20 percent lower than von Zahn's adjusted value, and is in agreement with values given by Tausch and Carignan (1972).

The concentration of helium in the lower thermosphere at midlatitudes is known to vary by a factor of as much as 10 between summer and winter. Also below 150 km, it appears not to be in diffusive equilibrium. The values presented in the present report fall between the extremes obtained in observations.

While some of the variations reported in $n(\text{N}_2)$, $n(\text{O}_2)$, and $n(\text{Ar})$ measurements in the 100- to 200-km range are almost certainly due to errors in measurements, some must be attributed to true atmospheric variations. The adopted values of $n(\text{N}_2)$, $n(\text{O}_2)$, and $n(\text{Ar})$ at 150 km listed in table 2 and used in constructing the present model are nominal values, and are the best estimates available at the present time. Each is believed to be corrected to 25 percent. Because of the uncertainty in the amount of atomic oxygen lost in rocket-borne mass spectrometers, the value of $n(\text{O})$ at 150 km is based on two sources. The first of these is the set of data obtained from the downward extrapolation of measurements made at higher altitudes with satellite-borne instruments in which, it is believed, the atomic-oxygen loss can be properly evaluated (Hedin et al., 1973). The second source is the set of mass densities found from satellite drag, and corrected for the other constituents (N_2 , O_2 , and Ar), which can be measured accurately. It appears likely that the $n(\text{O})$ values given in table 2 are maximum values, as they are based on the assumption that atomic oxygen is strongly absorbed in mass spectrometers used in rocket studies of the lower thermosphere. This view may be too pessimistic, but it does not seem probable that values given could be high by a factor as large as two.

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APPENDIX A

ATOMIC HYDROGEN

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For a standard atmosphere meant to represent average conditions, the thermospheric distribution of atomic hydrogen is considered to be adequately described by the steady-state equation for a minor constituent diffusing through a single stationary background gas:

$$\phi + D(H) \left[\frac{dn(H)}{dZ} + \frac{[1 + \alpha(H)] n(H)}{T} \cdot \frac{dT}{dZ} + \frac{n(H)}{H(H)} \right] = 0 \quad (25)$$

where

- $n(H)$ = number density of atomic hydrogen (m^{-3}),
- $H(H)$ = scale height of atomic hydrogen (m),
- $\alpha(H)$ = thermal-diffusion factor for atomic hydrogen, -0.25 (dimensionless),
- ϕ = vertical flux of atomic hydrogen ($m^{-2} s^{-1}$),
- $D(H)$ = mutual diffusion coefficient for atomic hydrogen ($m^2 s^{-1}$),
- T = temperature (K),
- Z = altitude (m).

One solution to equation 25 is

$$n(H) = \left[n(H)_r - \int_{Z_r}^Z \frac{\phi}{D(H)} \left(\frac{T}{T_r} \right)^{1+\alpha(H)} \cdot (\exp \tau) \cdot dZ \right] \left(\frac{T_r}{T} \right)^{1+\alpha(H)} \cdot (\exp - \tau) \quad (26)$$

where τ is the reduced height for atomic hydrogen defined by

$$\tau = \int_{Z_r}^Z \left(\frac{dZ}{H(H)} \right) \quad (27)$$

and the r subscripts refer to values at a reference level, Z_r . Equation 26 conveniently expresses the number-density distribution as the diffusive-equilibrium solution plus a contribution due to the flux ϕ . Equation 26 is used to compute the atomic hydrogen densities in the tables.

The nonzero flux for atomic hydrogen arises due to planetary escape from the exosphere, and the production of atomic hydrogen, H, from the reactions



in the lower thermosphere (Patterson, 1966). Due to the lack of knowledge regarding the effect of reactions 28 and 29, only that contribution to ϕ due to planetary escape from the exosphere is considered in the computations. This situation limits the validity of these computations to heights of 150 km and above. A constant value of $7.2 \times 10^{11} \text{ m}^{-2} \text{ s}^{-1}$ for the vertical flux is chosen as a compromise between the classical Jeans' escape flux for $T_\infty = 1000 \text{ K}$, with corrections to take into account deviations from a Maxwellian velocity distribution at the critical level (Brinkman, 1971), and the effects of charge exchange with H^+ and O^+ in the plasmasphere (Tinsley, 1973).

At a reference height of $Z_r = 500 \text{ km}$, the number density of atomic hydrogen, $n(\text{H})_r$, is taken to be $8.0 \times 10^{10} \text{ m}^{-3}$. This value is consistent with satellite data (Meier and Mange, 1970; Vidal-Madjar et al., 1973; and Brinton and Mayr, 1971 and 1972) appropriate for an exospheric temperature of 1000 K, and is approximately three times the value given in the early work of Kockarts and Nicolet (1963). The mutual diffusion coefficient $D(\text{H})$ is represented by the general expression

$$D_i = \frac{a_i}{N} \left[\frac{T}{273.15} \right]^{b_i} \quad (30)$$

where

$$a_i = 3.305 \times 10^{21} \text{ m}^{-1} \text{ s}^1,$$

$$b_i = 0.5 \text{ dimensionless, and}$$

$$N = \text{total number density of all significant species.}$$

This form for D_i arises from the kinetic-theory hard-sphere model for a two-component gas. The constant a_i represents an average between those for diffusion of H in N_2 , and for diffusion of H in O.

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APPENDIX B

THE 1976 STANDARD ATMOSPHERE AND ITS RELATIONSHIP TO EARLIER STANDARDS

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INTRODUCTION

A new U.S. Standard Atmosphere has been adopted (COESA, 1976). The definitions of this new Standard, along with graphs and abbreviated tables of the related atmospheric properties, have been published by Minzner et al. (1976). This new Standard was prepared in a format similar to that of the *U.S. Standard Atmosphere, 1962* (COESA, 1962), but differs in detail at all heights above 51 km.

This discussion examines the temperature-height profile and the corresponding density-height profile for the new Standard in relation to temperature-height and density-height profiles of all previous U.S. Standard Atmospheres. Also discussed is the history of the development of these standard atmospheres, and the relationship of these developments to the history of European standard atmospheres, beginning with an early isothermal atmosphere used primarily for altimetry purposes. This history concludes with the multi-purpose 1976 Standard Atmosphere which was designed to represent mean midlatitude conditions as presently known for heights within and below the thermosphere, and to represent mean solar conditions in and below the exosphere, where temperature is now known to vary in relation to the 11-year cycle of solar activity. The graphs presented here depict the evolution of our knowledge of the mean temperature and mean density of the Earth's atmosphere.

EARLY EUROPEAN STANDARDS

More than a century ago, aneroid barometers were calibrated for use as height-measuring instruments in accordance with an isothermal standard atmosphere. A British standard atmosphere of this era (Airy, 1867) provided a table of aneroid pressures as a function of height from 0 to 3657.6 m (12,000 ft) computed on the assumptions of a sea-level pressure of 31.0 in. of mercury and a constant mean temperature of 50°F (10°C), as depicted by line A₁ A₂ in figure 31.* This model was used as the basis of barometric altimetry in the United States until after World War I.

*The height coordinate in figure 31 is given in terms of geopotential kilometers, which will be discussed later.

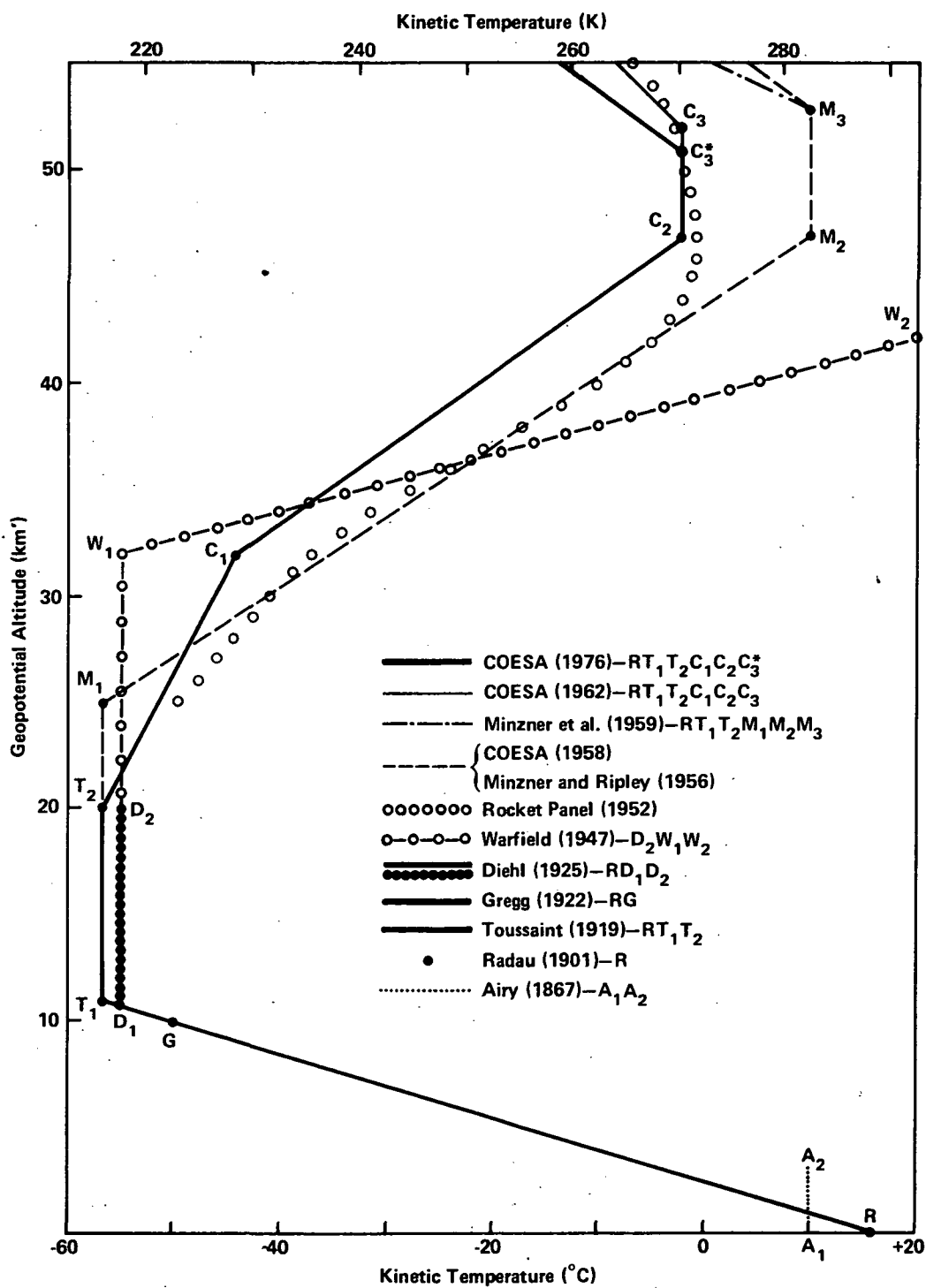


Figure 31. Kinetic temperature versus geopotential altitude.

During the early part of the twentieth century, the French had a different standard atmosphere for height determination, based upon a nonlinear temperature-height function proposed by Radau (1901). According to Grimault (1920), the French aeronautical community, until 1920, depended upon this "Law of Radau" as the temperature-height function in a standard atmosphere used for normalizing measurements of aircraft performance. This "law" expresses temperature as a function of pressure, with temperature decreasing asymptotically with increasing height from a constant value of 15°C at sea level (point R in figure 31), while pressure decreases from a constant sea-level value of 760 mm of mercury. These two sea-level constants were adopted into early international standard atmospheres, and have persisted in both international and U.S. standard atmospheres to the present time. The height-dependent portion of the "Law of Radau," however, was not retained in atmospheric models after 1920, since his expression was belatedly recognized to yield temperatures which, at heights above 5 km, departed significantly from existing measured values. These were average temperature-height profiles to altitudes of 14 km which had been computed for each of four seasons using observations obtained from balloon soundings in the region of Paris during a five-year period around the turn of the century (Teisserenc de Bort, 1904). At the time of publication, however, these profiles apparently had little impact upon the continued use of the Radau model of temperature versus height as the basis for barometric altimetry tables.

PROPOSAL FOR FIRST INTERNATIONAL STANDARD

The discrepancy between Radau's temperatures and those of the actual atmosphere appears to have gone unnoticed until the end of World War I, when Gamba (1918), in an Italian military report, published the graph of the 10-year average (1906 through 1916) of temperature versus height from 0 to 20 km for the region of Pavie, Italy. Using Gamba's curve as a basis for comparison, Toussaint (1919) noted the limitations of the Radau temperature-height expression, and showed the close agreement between Gamba's observations and a two-segment linear temperature-height function consisting of a constant negative gradient, $-0.0065^{\circ}\text{C/m}$ from $+15^{\circ}\text{C}$ at sea level to -56.6°C at 11,000-m altitude, and a zero gradient from 11,000- to 20,000-m altitude.

On the basis of this temperature-height function (RT_1 and $T_1 T_2$ in figure 31), which he proposed as the basis for an interallied aeronautical standard atmosphere, Toussaint computed tables of atmospheric pressure, temperature, and density as a function of altitude in increments of 500 m from 0 to 11,000 m, and in increments of 1000 m from 10,000 to 20,000 m. This proposal led the French (Grimault, 1920), in April 1920, to adopt the Toussaint temperature model as the basis of a new aeronautical standard atmosphere. Grimault, noting that the available meteorological data (primarily for western Europe) generally supported the Toussaint model, emphasized the advantages to be gained if cooperating European countries would each adopt the same standard. Italy and England soon followed this advice (Report of the British Aeronautical Research Committee, 1922), while the United States, apparently intending to do likewise, adopted only a portion of that model, a situation which led to confusion shortly thereafter and to some disagreement many years later.

FIRST U.S. AERONAUTICAL STANDARD

Prior to the French adoption of the Toussaint model, Gregg (1920) of the U.S. Weather Bureau, commenting on a competing European proposal by Soreau (1919), urged the rejection of the Soreau proposal and the adoption of the constant "rate of temperature decrease (of the Toussaint model)* resulting in values (which)* agree quite well with annual means as published by various investigators for Europe and the United States." In this document, Gregg also provided a table of computed pressures and densities for altitudes from 0 to only 10,000 m in accordance with the constant negative gradient of temperature $-0.0065^{\circ}\text{C/m}$ from a sea-level value of 15°C (shown as line RG in figure 31). He referred to this combination of gradient and sea-level value as Toussaint's "formula for computing 'standard' temperature for various heights."

In order to establish the basis for a U.S. Standard Atmosphere, Gregg (1922) said that the National Advisory Committee for Aeronautics (NACA) "requested the U.S. Weather Bureau to prepare a technical report covering the actual observations on the variation of temperature, pressure and density of the atmosphere for summer, winter and the year." In the conclusion of his report, Gregg stated that "It has been shown from observations over a long period that up to 10 or 12 km the mean variation of temperature with altitude in the United States is expressed very closely by Toussaint's formula." Again in this document, the observed atmospheric densities were compared with a table of computed densities (based upon the Toussaint model) which extended only from 0 to 10,000-m altitude (33,000 ft in the English tables), while the average observed values were tabulated up to 20,000-m altitude. This document also contained a table of other computed atmospheric parameters, but these too were limited to the height range of 0 to 10,000 m as in his earlier paper. Gregg appears to have known only about that lower portion of Toussaint's model involving the constant lapse rate of $-0.0065^{\circ}\text{C/m}$, and not the full 1919 version of that model.

The agreement between computed and observed U.S. data in the limited altitude range of 0 to 10,000 m appears to have been the basis for the adoption of the first U.S. Standard Atmosphere, for "At its meeting of December 17, 1921 the executive committee of the National Advisory Committee for Aeronautics (NACA) adopted for performance testing Toussaint's formula of temperature decrease with height for obtaining air density at different altitudes." (Gregg, 1922). This statement, with no height boundaries and no mention of an isothermal region, appears in a document which limited the listing of computed atmospheric properties to heights between 0 to 10,000 m while showing observed properties for a much greater height range. It implicitly limited the maximum height of the 1922 U.S. Standard Atmosphere to about 10,000 m (33,000 ft).

*Material in parentheses is not part of the original quotation.

In this *Standard Atmosphere* report, Gregg (1922) provided some direct translation of statements which he attributed to Toussaint regarding the uncertainty of the onset of an isothermal layer in European regions, but cited no specific document as a reference for these remarks. In this same report, Gregg also stated that "Toussaint has not carried his computations above 10 kilometers." Obviously, Gregg must have been dealing with some Toussaint communication which predated the Toussaint publication of October 1919, for the latter document certainly did specify the onset of the isothermal layer to occur at 11,000 m and did contain a table of computed atmospheric properties for heights from 11,000 to 20,000 m, as well as one for heights from 0 to 11,000 m. Thus, while strongly recommending that the United States officially adopt the Toussaint model, and while recognizing that France, Italy, and England had already done so, Gregg apparently was not aware of the complete details of the models which these countries had adopted.

In particular, neither Gregg nor the Executive Committee of NACA appear to have been aware of NACA Technical Memorandum No. 15 (NACA, 1921) prepared by the Paris office of the NACA prior to April 1921. This Memorandum consisted of a translation of an expanded version of the paper by Grimault (1920) in which the adoption of the Toussaint model by the French government was discussed, along with detailed tables and formulas for computing temperature, pressure, and density in each of two height regions—0 to 11,000 m and 11,000 to 20,000 m. Supporting observational data from various sources were also republished in this report.

The 1921 action of the Executive Committee of NACA was not in conflict with a hoped-for interallied standard atmosphere, but this action did allow for ambiguity which led to a disagreement between subsequent U.S. Standard Atmospheres and the standard atmosphere of the International Commission for Air Navigation (ICAN), which was adopted in 1924 (ICAN, 1924), and subsequently clarified (ICAN, 1925, 1938, 1951) on the basis of the complete Toussaint model.

EXPANSION OF THE FIRST U.S. STANDARD ATMOSPHERE

The *Standard Atmosphere* report prepared by Gregg (1922) dealt primarily with the relationship of meteorological data to the tabulated values of the lower portion of the atmospheric model of Toussaint. Diehl (1922), concerned with the engineering aspects of a model atmosphere, investigated the mathematical relationships between various atmospheric properties for the case when temperature decreases linearly with increasing altitude. The growing requirements for standard atmosphere tables by the aeronautical community as well as by artillerists led Diehl (1925) to expand the U.S. Standard Atmosphere. This expansion included the following:

- Increasing the height resolution,
- Adding listings of ratios of the several properties to their sea-level values, and
- Extending the altitude range from a maximum of 10,000 m up to 20,000 m (but in a manner not consistent with the Toussaint model in the ICAN Standard).

Diehl, like Gregg, appears to have been unacquainted with the complete Toussaint model as well as with the actions of ICAN, for while he properly extended the range of the constant lapse rate of temperature ($-0.0065^{\circ}\text{C}/\text{m}$ used by Gregg) to heights above 10,000 m, he unfortunately extended it only to a height of 10,769 m (see GD_1 in figure 31), where the computed temperature becomes -55°C (in accordance with Gregg's mean of observed temperatures for 45°N in the United States). Then, from 10,769- to 20,000-m altitude, Diehl introduced an isothermal layer at -55°C (see $D_1 D_2$ in figure 31) instead of the one previously adopted at -56.5°C in the Toussaint-ICAN standard ($T_1 T_2$ in figure 31). The slight difference in the height and temperature of the tropopause (231 m and 1.5°C , respectively) later led to international aeronautical confusion which was not resolved until after World War II.

In the interim it was decided that, in the United States, in accordance with the actions of ICAN (ICAN, 1924), altimetry would no longer be based upon an atmosphere isothermal at 10°C (NBS, 1920), but rather would be based on the same standard used for comparison of aircraft performance. The metric altimetry tables of ICAN, however, were neither in agreement with the U.S. Standard Atmosphere (Diehl, 1925) at heights above 10,769 m, nor convenient in the United States where measurements were made primarily in English units. Furthermore, Diehl's English tables of pressure versus height were not conveniently formulated for altimeter calibration. Consequently, Brombacher (1926, 1935) of the U.S. Bureau of Standards prepared tables of altitude as a function of pressure as well as other tables particularly applicable to the calibration of altimeters.

Upper Air Extension of the U.S. Standard Atmosphere

The Diehl and Brombacher tables adequately served the aeronautically isolated United States until, during, and after World War II when the development of high-altitude weapons required the use of tables of atmospheric properties to heights far in excess of the 20-km limit of both the U.S. Standard and of the Toussaint-ICAN Standard. In the United States, this situation was remedied by Warfield (1947), who prepared *Tentative Tables for the Properties of the Upper Atmosphere* extending from the 20-km top of the Diehl Standard up to 120-km altitude. These tables were based upon temperatures measured to 32 km by balloons, and on indirectly deduced values of temperature between 32 and 120 km. As in the Diehl standard, Warfield's temperature-height profile consisted of a series of linear segments ($D_2 W_1$ and $W_1 W_2$ in figure 31, plus $W_2 W_3$, and so on in figure 32). The lowest segment ($D_2 W_1$), from 20- to 32-km altitude, represents an isothermal extension of the isothermal layer which had been defined earlier by Diehl for the height region from 10.769 to 20 km.

Warfield recognized the need for considering the effect of the height dependence of the acceleration of gravity on atmospheric properties, particularly at high altitudes. He consequently computed two sets of tables: one of these sets assumed an inverse square-law variation for the acceleration of gravity with height, while the other assumed the acceleration of gravity to be invariant with height, in accordance with previous custom. This assumption is tantamount to replacing geometric height, measured in geometric kilometers

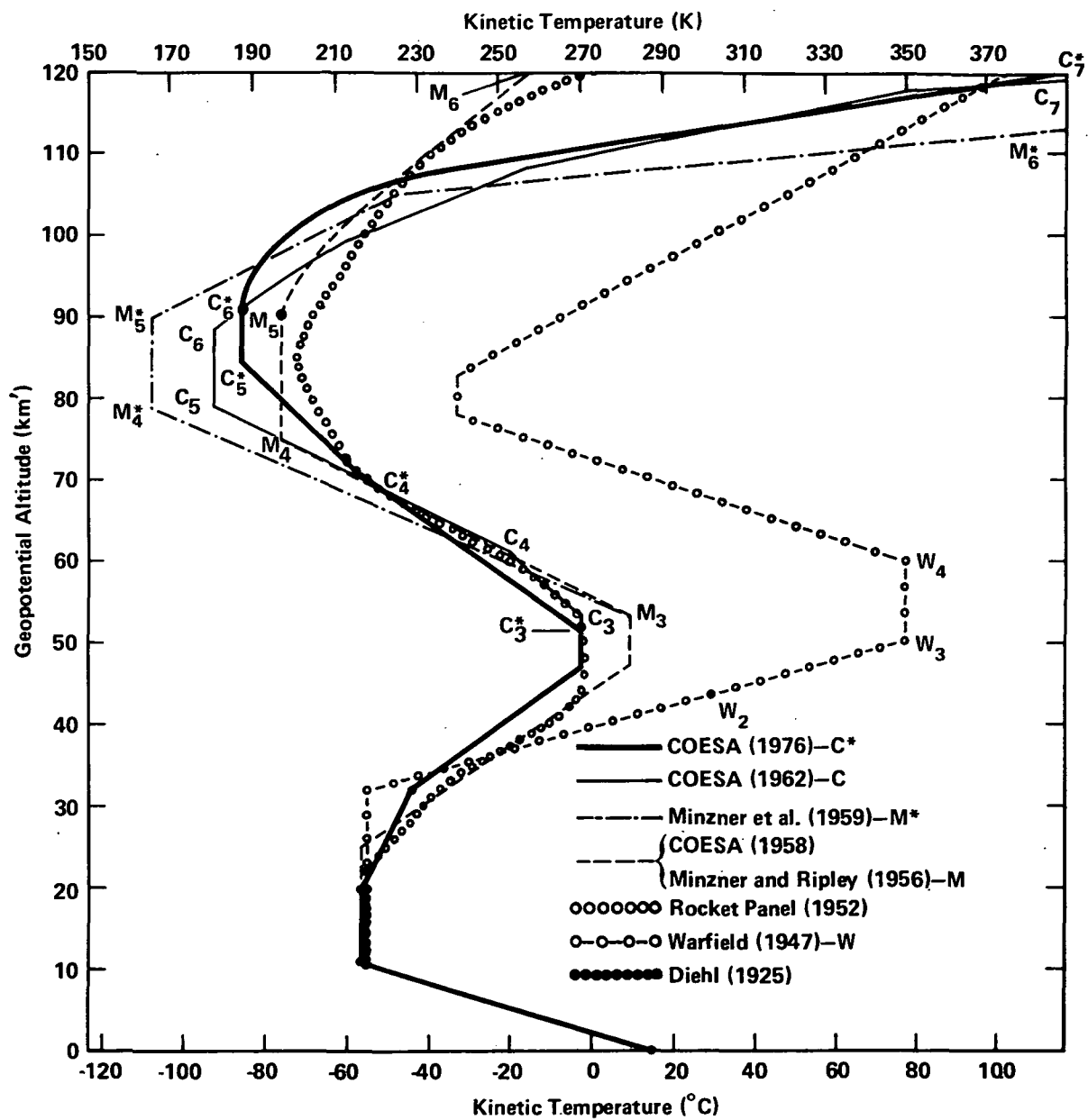


Figure 32. Kinetic temperature versus geopotential altitude.

(km), with geopotential height, measured in geopotential kilometers (km'), the argument upon which the current International Standard Atmosphere is based. The height coordinate in figure 31, as well as in figure 32, is therefore given in terms of geopotential kilometers.

NEW ATTEMPTS AT THE DEVELOPMENT OF AN INTERNATIONAL STANDARD ATMOSPHERE

Prior to World War II, the normal aircraft flight altitudes were usually limited to heights below 10 km. The war-time developments, however, were about to raise the cruising altitude of commercial aircraft to heights above 10 km, where the basis for the U.S. Standard differed from the several versions of the ICAN Standard generally used in Europe (Brombacher, 1944). Thus, while one group at NACA was concerned with extending the height range of the current U.S. Standard Atmosphere, another group at NACA became concerned with the elimination of the differences between the U.S. Standard Atmosphere and the standard atmospheres used by other countries. The continuation of the cooperation developed between the United States and its allies during World War II made this period an ideal one for the latter task.

A series of meetings of the International Civil Aviation Organization (ICAO) led to a compromise (ICAO, 1950): the United States agreed to the adoption of the entire Toussaint temperature-height profile (RT_1 and $T_1 T_2$ in figure 31), which had long been a part of the ICAN Standard, while the former users of the ICAN Standard agreed to the adoption of values of certain nominal constants, such as the sea-level pressure and the sea-level acceleration of gravity at 45°N in accordance with current usage in the United States. The ICAO also agreed to the use of certain innovations, such as calculations of pressure and density on the basis of a height-dependent acceleration of gravity, when these properties were to be tabulated as a function of geometric height.

These compromises led to the ICAO Standard Atmosphere, the definitions of which were adopted for the United States by NACA on November 20, 1952. Tables based upon these definitions were computed jointly by Italy and the United States, and the U.S. version of these tables (NACA, 1954, 1955) represents the first set of standard atmosphere tables produced entirely by automatic computing machines.

INITIAL U.S. ATTEMPT AT EXTENDING THE ICAO STANDARD

This new international standard atmosphere adopted by the United States included the height range of -5,000 to only 20,000 m, but it essentially eliminated the further use of the Warfield upper-atmosphere tables, which were not continuous with the new Standard, and which by this time were recognized to involve excessively high temperatures at heights above 40 km. The high-altitude rocket development and atmospheric research being pursued throughout the world by this time, however, indicated a definite need for an extension of the ICAO Standard Atmosphere to much greater heights than its 20-km limit. For purposes of atmospheric modeling, the Rocket Panel (1952) had independently adopted a

digitally defined temperature-height profile from 0 to 220 km. For heights of 25 km and above, these temperatures (shown in figures 31, 32, and 33 as a line of small circles) were inferred from a small number of rocket soundings involving a number of different measurement techniques. (Note the change in the units of height from geopotential kilometers (km') in figures 31 and 32 to geometric kilometers (km) in figure 33. This is because the later and higher models had their upper portions defined in terms of geometric rather than geopotential kilometers. This was also the situation for the entire Rocket Panel model.)

The availability of an ever-increasing amount of atmospheric sounding data at heights up to 120 km, coupled with the increasing need for a high-altitude standard atmosphere, led the U.S. Weather Bureau and the Geophysics Research Directorate of the Air Force Cambridge Research Center (AFCRC) to cosponsor the formation of the Committee on Extension to the Standard Atmosphere (COESA). At its first meeting in November 1953, this committee assumed the responsibility of reviewing existing data regarding the upper atmosphere, and preparing a U.S. extension to the ICAO Standard Atmosphere to altitudes above 20 km.

The Working Group of COESA recommended a particular linearly segmented function of molecular-scale temperature (Minzner and Ripley, 1956) versus height from 20- to 300-km' altitude. An extended version of the function was immediately adopted as the basis of an engineering model atmosphere for use by the Air Research and Development Command (ARDC) of the U.S. Air Force, and hence, this model, which extended to an altitude of 542.7 km (500 km') became known as the *ARDC Model Atmosphere, 1956* (Minzner and Ripley, 1956). The defining height profile of molecular-scale temperature in this model began with the complete Toussaint model and continued with an extension of Toussaint's isothermal layer at -56.5°C from 20 to 25 km'. For heights above 20 km', this temperature function is represented by the linear segments $T_2 M_1$, $M_1 M_2$, and $M_2 M_3$ in figure 31; it continues upward with segments $M_3 M_4$, $M_4 M_5$, and $M_5 M_6$ in figure 32; and finally extends upward through M_7 to M_8 in figure 33. (When this and other models defined in terms of linear segments of molecular-scale temperature versus height are depicted in terms of kinetic temperature versus height, the kinetic temperature segments are usually not linear at heights above 90 km, where mean molecular weight begins to decrease with increasing height.)

That portion of this same temperature-height function up to 300 km' (314.9 km), designated by point M_7 in figure 33, was adopted by COESA (1956) as the basis for the *U.S. Extension to the ICAO Standard Atmosphere* (COESA, 1958). For this extension, ICAO agreed that the height region from -5 to 32 km' was to be designated as standard, with the suggestion that the particular portion of this model from 20- to 32-km' altitude should be adopted by ICAO as an extension to its Standard. The height region from 32 to 75 km' was to be designated as tentative, while that from 75 to 300 km' was to be speculative. These latter two designations were adopted to allow for future changes as our knowledge of the atmosphere was refined. Other recommendations adopted by COESA concerning this model included rather peculiar height increments in the metric tables. These height increments of 150, 300, and 750 m in successive height regions of the metric tables (to approximate the increments of 500, 1000, and 5000 ft in corresponding height regions of

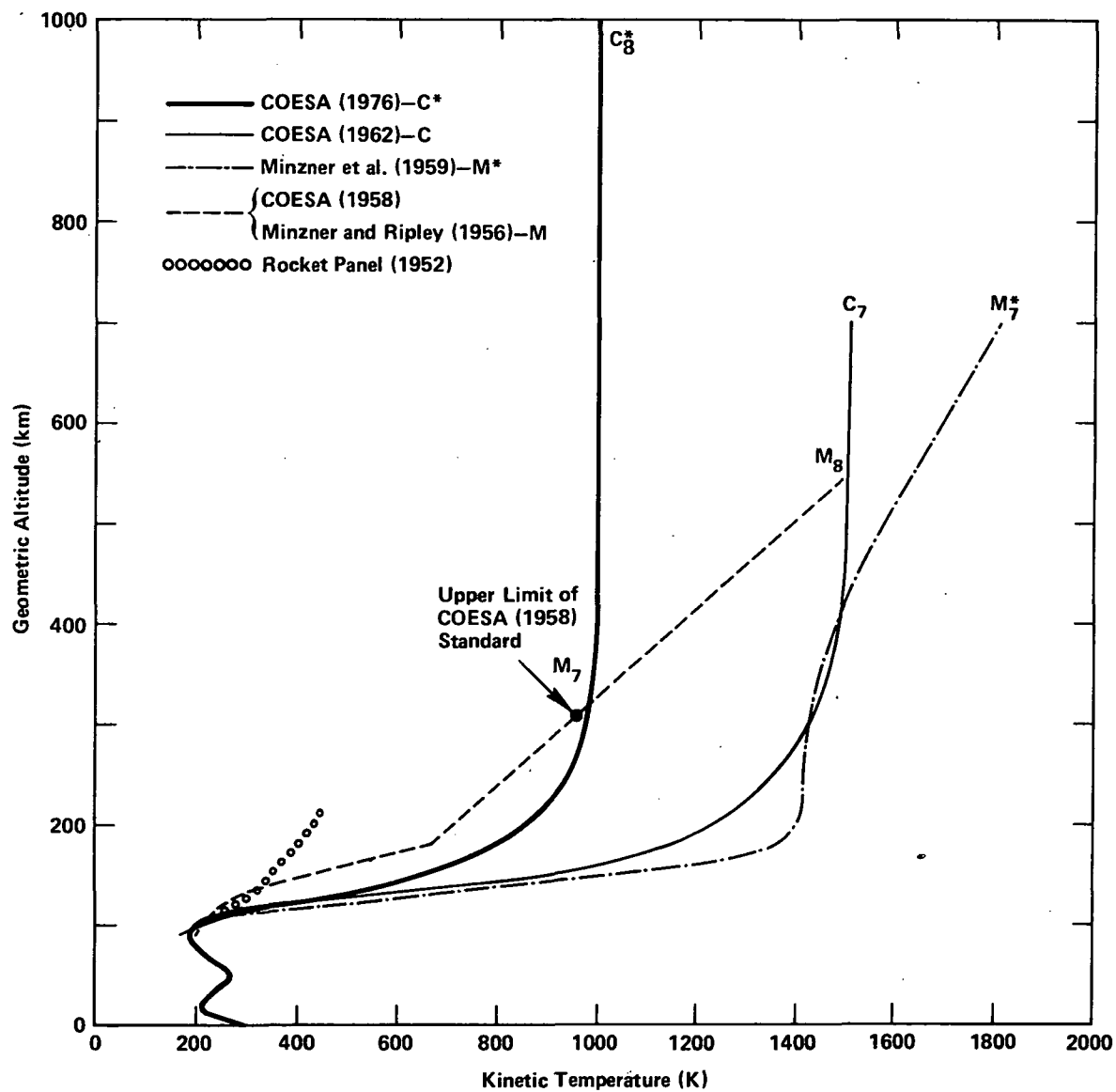


Figure 33. Kinetic temperature (with varying mean molecular height) versus geometric altitude.

the English tables) made it impossible for metric heights to be given in successive integer multiples of 1 km. Formal tables based upon these peculiar height specifications were prepared by Minzner et al. (1958), but these tables were found to be inconvenient for people working in the metric system. The metric portion of the earlier tables of this model in the form of the *ARDC Model Atmosphere* (Minzner and Ripley, 1956), while prepared with the coarser height resolutions of 1 km from -5 to 132 km, and of 5 km from 135 to 500 km, appeared to be more convenient than the metric portion of the formal tables of the 1958 *U.S. Extension to the ICAO Standard Atmosphere* (Minzner et al., 1958).

The 1959 High-altitude Revision

The launching of the Sputnik-I satellite by the Russians on October 4, 1957 introduced a new era in space research. The analysis of orbital decay characteristics of this and seven other satellites (through Explorer-4), which were launched during the next 22 months, led to several dozen values of atmospheric density at altitudes between 160 and 600 km. These densities indicated the need to reexamine the 1956 ARDC Model Atmosphere, and the related 1958 U.S. Standard Atmosphere at all heights above 100 km. The atmospheric densities inferred from the observed decay of satellite orbits were greater than those of the 1956 Model by factors ranging from about 3 at 200-km altitude, through about 10 at 300-km, to more than 18 at the 500-km level, thereby implying the need to change the defining temperature-height profile of the model. Simultaneously, new rocket sounding data indicated the need to decrease the densities of the model in the height region between 85 and 145 km.

A revised altitude profile of kinetic temperature for heights above 53 km (M_3 , M_4^* , $M_4^*M_5^*$, and so on through M_6^* in figure 32, and to M_7^* in figure 33) was based upon a specified set of linear segments of molecular-scale temperature versus geopotential height. The successive members of the set of linear segments were determined iteratively to yield computed densities which were in agreement with the mean of those derived from rocket soundings and from satellite drag data. This set of successive linear functions defined the *ARDC Model Atmosphere, 1959* (Minzner et al., 1959). Below 53-km altitude, this model was identical to both the *ARDC Model Atmosphere, 1956* (Minzner and Ripley, 1956) and the related *U.S. Extension to the ICAO Standard Atmosphere* (Minzner et al., 1958). The 1959 Model differed from its predecessors in two significant ways. One change was the great increase in density, particularly at heights above 200 km, as shown in figure 34. The other change was the extension of the upper limit of the model from 500 to 700 km, as shown in figures 33 and 34. The 1959 Model also represented an improvement in densities between 100 and 120 km.

ICAO Rejection of Initial Attempt at Extension

In international circles, the *U.S. Extension to the ICAO Standard Atmosphere* (Minzner et al., 1958), upon which the *ARDC Model Atmosphere, 1959* (Minzner et al., 1959) was based, was not well received. This was because the upward continuation of the Toussaint

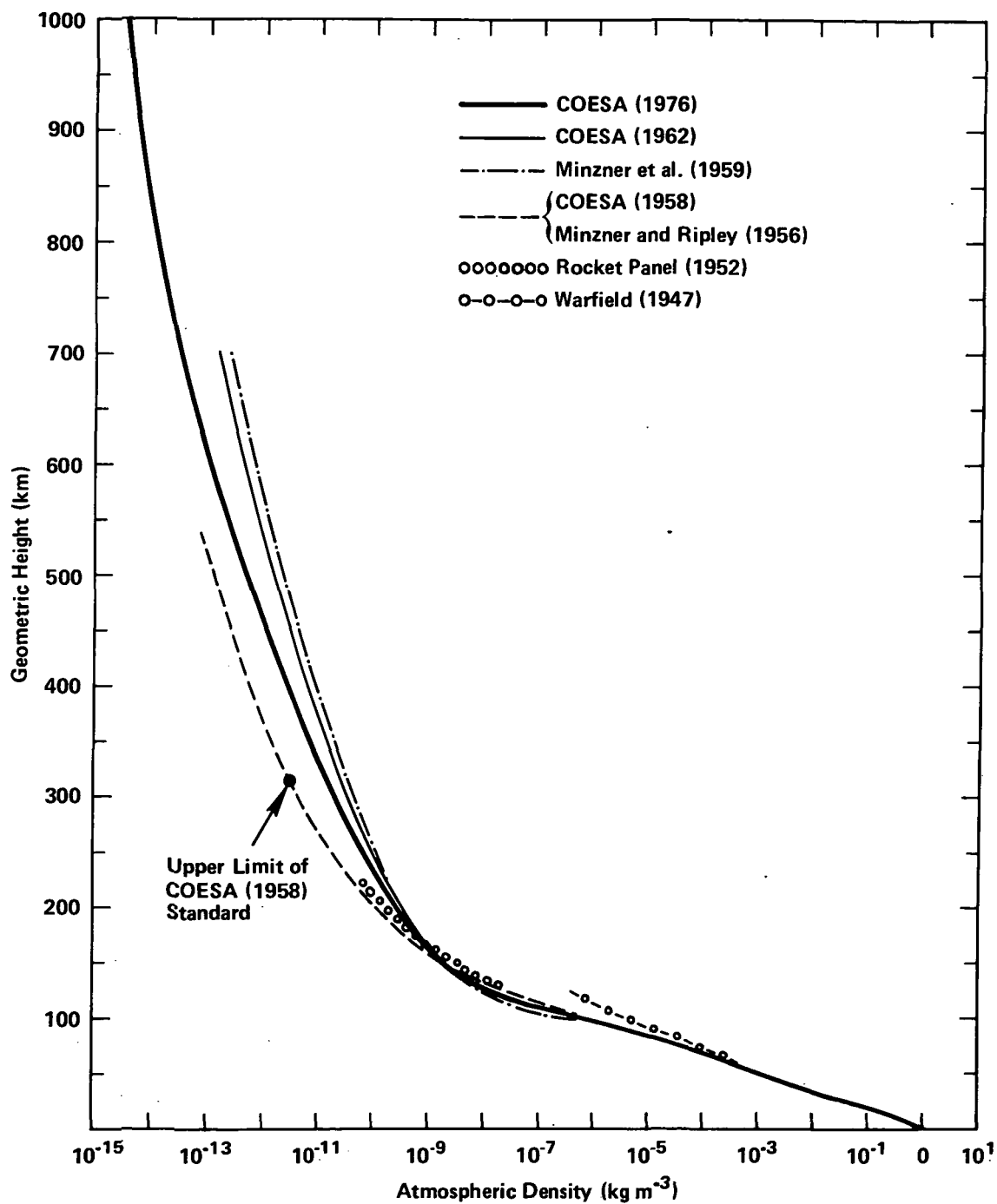


Figure 34. Atmospheric density versus geometric height for the 1976 U.S. Standard Atmosphere and for four earlier U.S. standard and model atmospheres.

isothermal layer from 20 to 25 km in the 1958 extension was not in agreement with the mean observed temperature of this height region (which by this time was being very well measured). This lack of agreement with observations had been well known to the agencies involved in the preparation of the *U.S. Extension to the ICAO Standard Atmosphere* (see Rocket Panel temperatures in figure 31), but because of some vested interests (it was later learned) involving very expensive equipment already built and calibrated according to such a temperature-height profile, some of the U.S. agencies of COESA held out for this unrealistically extended isothermal layer. ICAO, however, refused to adopt any portion of this 1958 *U.S. Extension to the ICAO Standard Atmosphere* as a basis for extending its own international standard. COESA was forced to review its position.

A Second Attempt at Extending the ICAO Standard

The Working Group of COESA reconvened in 1960, and through 1962, reviewed the ever-increasing set of high-altitude atmospheric density data being inferred from satellite orbits, as well as the large set of density data inferred from rocket soundings. The number of rocket soundings had increased greatly under the stimulus of the International Geophysical Year (IGY) which extended over the 18-month period from July 1957 through December 1958. This review led to recommendations of a linearly segmented temperature-height profile which was significantly changed from that of the 1958 *U.S. Extension to the ICAO Standard Atmosphere* (Minzner et al., 1958), with changes beginning at heights as low as 20 km, in keeping with the desires of the ICAO nations, and extending to a height of 700 km, in keeping with temperatures inferred from the now much larger sets of rocket soundings and satellite orbits. This recommended model, which was developed during 1962, was defined by a series of 19 linear functions of molecular-scale temperature versus height. The first six of these, extending over the 20- to 88.743-km' (90-km) height region, are linear with respect to geopotential. The remaining 13 functions (which specify the 90- to 700-km region) were defined to be linear with respect to geometric height in deference to aeronomers and space scientists who are among the primary users of atmospheric models in this height region, but who rarely deal with the concept of geopotential.

The first three of the functions which are linear with respect to geopotential are depicted in figure 31 by segments $T_2 C_1$, $C_1 C_2$, and $C_2 C_3$. They have gradients of +1.0, +2.8, and $0.0^\circ\text{C}/\text{km}'$, respectively. The remaining three functions which are linear with respect to geopotential, are depicted in figure 32 by segments $C_3 C_4$, $C_4 C_5$, and $C_5 C_6$. The 13 functions which are linear with respect to geometric height are not specifically designated, but extend from C_6 in figure 32 through C_7 in figure 33.

The segment $T_2 C_1$ in figure 31, with its gradient of $+1.0^\circ\text{C}/\text{km}'$, was selected for the 1962 Model to replace segments $T_2 M_1$ and a portion of segment $M_1 M_2$ of the 1958 Standard for the 20- to 32-km' height region because this segment provided a better fit to the known temperature data, and consequently, would have a better chance of acceptance by ICAO. Portions of the temperature-height function of the 1958 U.S. Standard for heights above 32 km' were changed to the new values (designated by $C_1 C_2$ and $C_2 C_3$ in figure 31), not

only to approach observed temperature data more closely, but also to compensate for the effect which the changes below 32 km' had upon the computed densities of the 1962 Model for heights to 50 km' and above.

Densities determined from satellite drag acceleration measured between 1958 and 1960 suggested the need for a reduction in the densities of the 1959 ARDC Model Atmosphere at heights above 200 km since these 1959 densities appeared to be peak values associated with a maximum in solar activity. Consequently, the temperatures of the 1962 Model were adjusted to lead to a slight reduction in the computed densities for heights above 200 km from those of the 1959 Model as shown in figure 34.

On March 15, 1962, COESA (at this time representing 29 scientific and engineering organizations in the United States) adopted this 1962 Model as the *U.S. Standard Atmosphere, 1962* (COESA, 1962), which between 20 and 32 km was consistent with the desires of ICAO representatives. COESA again recommended that the ICAO Standard Atmosphere be extended from its 20-km altitude limit of 1954 (NACA, 1954) up to a height of 32 km following the definitions of this revised U.S. Standard. This recommendation was soon adopted by ICAO (1964).

SUPPLEMENTS TO THE 1962 U.S. STANDARD

The *U.S. Standard Atmosphere, 1962* (COESA, 1962) was designed to represent midlatitude annual mean conditions. The large amount of atmospheric data gathered during the IGY, however, permitted the generation of mean atmosphere models for various seasons and several latitude bands, as well as for various degrees of solar and geomagnetic activity. Consequently, COESA extended its interests to activities beyond the standard atmosphere by preparing the *U.S. Standard Atmosphere Supplements, 1966* (COESA, 1966) which contains two series of tables of atmospheric properties. The first series of tables is for the 0- to 118-km height range. It consists of separate sets of tables for January and July for each of four latitudes—30°N, 45°N, 60°N, and 75°N—as well as a set for mean annual conditions at 15°N and a set for spring-fall conditions at midlatitudes. (The tables for 75°N do not extend above 30-km altitude.) The second series of tables is for the 120- to 1000-km height range and consists of separate sets of tables for three seasonal periods (winter, summer, and a combined spring-fall period) for each of 11 exospheric temperatures between 600 and 2100 K.

THE U.S. STANDARD ATMOSPHERE, 1976

At the time of the preparation of the *U.S. Standard Atmosphere, 1962* (COESA, 1962), it was suggested that the value of the atmospheric density at heights above 200 km depended upon solar activity. This dependence was demonstrated in part by the work supporting the *U.S. Standard Atmosphere Supplements, 1966*. This dependence was demonstrated even more convincingly during the following years from analyses of high-altitude densities measured during more than one complete solar cycle (Jacchia, 1971). These measurements indicated that when densities are suitably averaged over one solar cycle, the mean density at any

particular height above 200 km should be considerably lower than indicated by the 1962 Standard.

The set of rocket soundings for the 50- to 120-km height region had also grown greatly during the 10-year period following the preparation of the 1962 Standard. These soundings indicated that significant adjustments were also needed in the density-height profile of the 1962 Standard at heights as low as 70 km.

In light of these situations, the COESA Working Group was reconvened in September 1971 to prepare recommendations for a revision to the *U.S. Standard Atmosphere, 1962* (COESA, 1962). Through the activities of three task groups working intermittently over the next three years, the compromises and recommendations leading to the definitions of the *U.S. Standard Atmosphere, 1976* (COESA, 1976 and Minzner et al., 1976) were laboriously generated. This new standard atmosphere, officially adopted by COESA in February 1975, is identical to the 1962 version of the U.S. Standard Atmosphere for heights up to and including 51 km. It involves moderate departures from the 1962 Standard for heights between 51 and 150 km, and extensive revisions for heights above 200 km. The densities of the 1976 Standard are about 10 percent lower than those of the 1962 Standard in each of two height regions—70 to 80 km, and 115 to 120 km, as shown in figure 35. This figure also shows the densities of the 1976 Standard increased by about 10 percent over the 1962 values in each of two other height regions—90 to 105 km, and 132 to 150 km. Above 165 km, the depression of the densities of the 1976 Standard, relative to those of the 1962 Standard, increase steadily with increasing height such that, at 700-km altitude, the density of the 1976 Standard is about one-seventh that of the 1962 Standard, as shown in figure 34.

The differences between the density-height profile of the 1962 Standard and that of the 1976 Standard are associated with related differences between the temperature-height profiles of the two standards. These temperature differences are evident in part from a comparison of the two temperature-height profiles in figures 31, 32, and 33 at all heights above 51 km. In figures 31 and 32, the 1976 Standard is seen to depart from the isothermal layer at a height 1-km' lower than the point where the 1962 Standard departs from the same isothermal. The segments $C_3^*C_4^*$, $C_4^*C_5^*$, and $C_5^*C_6^*$ shown in figure 32 are each linear with respect to geopotential, and represent gradients of -2.8, -2.0, and 0.0 K/km', respectively. These values, along with other constants, define the 1976 Standard at heights between 51 and 84.854 km' (86 km). (The 1976 Standard is defined in terms of kinetic temperature while the 1962 Standard was defined in terms of molecular-scale temperature. At heights below 86 km, the difference between kinetic temperature and molecular-scale temperature is essentially negligible while the difference may approach 10 percent as height increases from 86 to 120 km.)

At heights above 86 km, the kinetic temperature of the 1976 Standard is defined as a series of four consecutive functions of geometric height so chosen that the first derivative of temperature with respect to height is continuous over the entire height region of 86 to 1000 km (between point C_6^* in figure 32 and C_8^* in figure 33).

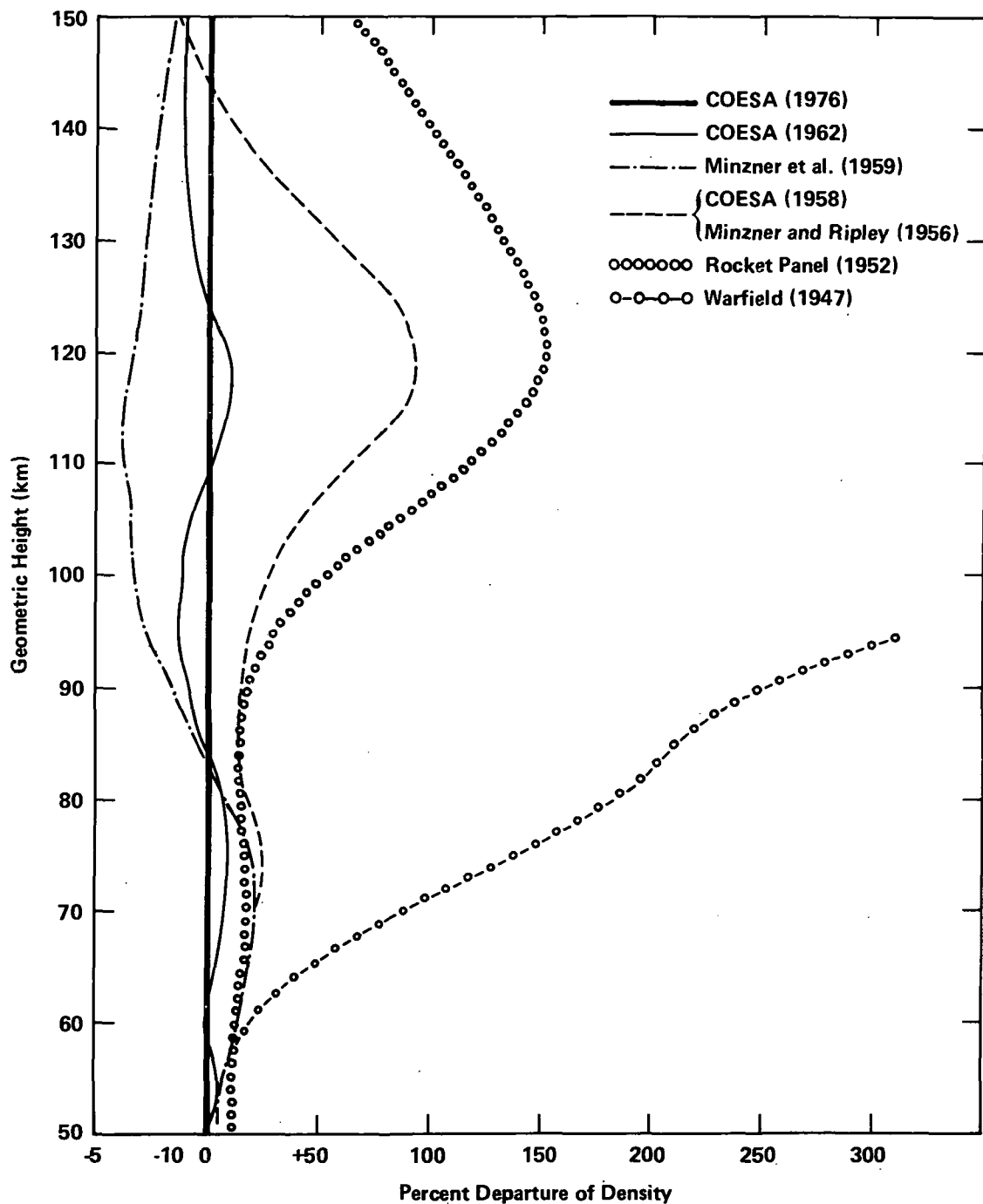


Figure 35. Percent departures of density of various previous standard and model atmospheres from the 1976 U.S. Standard Atmosphere.

The greatest difference in kinetic temperature between the two models is seen in figure 33 which shows a -500 K difference between the exospheric temperatures of the two standards extending downward to a height of about 400 km. This temperature difference decreases to 400 K, 300 K, and 200 K at altitudes of about 260, 160, and 140 km, respectively. The greatest height at which the kinetic temperatures of the two standards are equal is about 122 km. The temperature reduction above 122 km is primarily responsible for the decrease in the densities of the 1976 Standard above 150 km. The much smaller differences between the temperatures of the two models in the height region of 51 to 120 km', as seen in figure 32, are associated primarily with the smaller adjustments in the density-height profile in the same height region. However, the effect of any change in temperature at one altitude propagates upward in the density-height profile as a negative exponential function of the integral of the reciprocal of the temperature times the differential of height. Consequently, changes in the temperature-height profile below 120 km influence the entire density-height profile above this altitude. Furthermore, the relative effect of a 1° change over a 1-km height increment near the mesopause where the temperature is 187 K is much greater than the effect of a similar change at 120 km where the temperature is 360 K. Consequently, the shape of the temperature-height profile near the mesopause was found to be quite critical in achieving the desired density-height profile at heights as great as 200 km.

The processes of dissociation of molecular oxygen and of molecular diffusion, both occurring within the height region between 86 and 200 km, also cause the modeling of this height region to be more complex than that for other regions of the 1976 Standard. Furthermore, the observational data for this region are more meager and less reliable than those for other height regions. Consequently, the agreements and recommendations for the details of this height region in the 1976 U.S. Standard Atmosphere were generated only after considerable effort and successive revision, and may still be open to question.

The successive atmospheric models depicted in figures 31 through 35 suggest a process of gradual convergence toward the mean static model described by the *U.S. Standard Atmosphere, 1976* (COESA, 1976). While the Working Group of COESA believes this model to be a good representation of the currently known average atmospheric conditions, it is not known whether future observations will change that opinion. Barring long-term climatic changes, it appears unlikely that future data will suggest the need for significant changes in a mean static model atmosphere for 45°N latitude at heights below 80 km. For heights between 100 and 200 km, however, the possibility that future data will continue to support the 1976 Model is somewhat decreased. Any future changes in the mean atmospheric model in this height region would, of course, propagate the need for changes at greater altitudes.

The extensive range of experience, as well as the competence of the members of the four task groups involved in the generation of the 1976 Standard, suggests that this Standard can be expected to be a reasonable and lasting midlatitude static model.

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